

Meridional Heat Transport in the DeepMIP Eocene ensemble: non-CO2 and CO2 effects

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Abstract

The total meridional heat transport (MHT) is relatively stable across different climates. Nevertheless, the strength of individual processes contributing to the total transport are not stable. Here we investigate the MHT and its main components especially in the atmosphere, in five coupled climate model simulations from the Deep-Time Model Intercomparison Project (DeepMIP). These simulations target the Early Eocene Climatic Optimum (EECO), a geological time period with high CO₂ concentrations, analogous to the upper range of end-of-century CO₂ projections. Preindustrial and early Eocene simulations at a range of CO₂ levels (1x, 3x and 6x preindustrial values) are used to quantify the MHT changes in response to both CO₂ and non-CO₂ related forcings. We found that atmospheric poleward heat transport increases with CO₂, while the effect of non-CO₂ boundary conditions (e.g., paleogeography, land ice, vegetation) is causing more poleward atmospheric heat transport on the Northern and less on the Southern Hemisphere. The changes in paleogeography increase the heat transport via transient eddies at the mid-latitudes in the Eocene. The Hadley cells have an asymmetric response to both the CO₂ and non-CO₂ constraints. The poleward latent heat transport of monsoon systems increases with rising CO₂ concentrations, but this effect is offset by the Eocene topography. Our results show that the changes in the monsoon systems' latent heat transport is a robust feature of CO₂ warming, which is in line with the currently observed precipitation increase of present day monsoon systems.

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21 **Key Points:**

- 22 • The latent heat transport of the monsoon systems increases through higher CO₂
23 concentration, but it is reduced by the Eocene topography.
- 24 • The poleward heat transport of midlatitude cyclones is higher in the Northern
25 Hemisphere in the Eocene, due to the different topography.
- 26 • The Hadley cells are overturning more heat in response to the higher CO₂ values, but the
27 net poleward heat transport is relatively stable.

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29

30 **Abstract**

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 32 Nevertheless, the strength of individual processes contributing to the total transport are not
 33 stable. Here we investigate the MHT and its main components especially in the atmosphere, in
 34 five coupled climate model simulations from the Deep-Time Model Intercomparison Project
 35 (DeepMIP). These simulations target the Early Eocene Climatic Optimum (EECO), a geological
 36 time period with high CO₂ concentrations, analogous to the upper range of end-of-century CO₂
 37 projections. Preindustrial and early Eocene simulations at a range of CO₂ levels (1x, 3x and 6x
 38 preindustrial values) are used to quantify the MHT changes in response to both CO₂ and non-
 39 CO₂ related forcings. We found that atmospheric poleward heat transport increases with CO₂,
 40 while the effect of non-CO₂ boundary conditions (e.g., paleogeography, land ice, vegetation) is
 41 causing more poleward atmospheric heat transport on the Northern and less on the Southern
 42 Hemisphere. The changes in paleogeography increase the heat transport via transient eddies at
 43 the mid-latitudes in the Eocene. The Hadley cells have an asymmetric response to both the CO₂
 44 and non-CO₂ constraints. The poleward latent heat transport of monsoon systems increases with
 45 rising CO₂ concentrations, but this effect is offset by the Eocene topography. Our results show
 46 that the changes in the monsoon systems' latent heat transport is a robust feature of CO₂
 47 warming, which is in line with the currently observed precipitation increase of present day
 48 monsoon systems.

49

50 Plain Language Summary

51 In the Earth's climate system both the atmosphere and the ocean are transporting heat
 52 through different processes from the tropics towards the poles. We investigate the transport of
 53 the atmosphere in several climate model set ups, which are aiming to describe the very warm
 54 climate of the early Eocene (~56-48 Myr ago). This period is relevant for us, because the
 55 atmospheric CO₂ concentration was close to our pessimistic projection of CO₂ concentration for
 56 the end of the century. In our study we are separating the results depending on their origin,
 57 meaning if the changes, seen in the heat transport are due to the larger CO₂ concentration or due
 58 to the different set up of the Eocene. We found that with rising CO₂ values the atmosphere
 59 transports more heat from the tropics to the polar regions. The different location of the continents
 60 and seas is influencing the heat transport of the midlatitude cyclones. The monsoon systems
 61 seem to be affecting a globally smaller area in the Eocene, but being more effective in
 62 transporting heat due to the higher atmospheric CO₂ concentration. This conclusion is in line
 63 with the observation, that current day monsoon systems' precipitation increases, as our CO₂
 64 concentration rises.

65

66 **1 Introduction**

67 The meridional temperature gradient is the main driving force of the atmospheric and
 68 oceanic general circulation. It is caused by differential radiative heating and leads to meridional
 69 heat transport from the tropics to the mid- and high latitudes. It has been shown that the
 70 Meridional Heat Transport (MHT) is stable in different climate states (Krapp & Jungclaus, 2011;
 71 Smith et al., 2006; Yang et al., 2015). Bjerknes (1964) proposed that if the net radiation forcing
 72 at the top of the atmosphere (TOA) and the ocean heat storage do not vary too much, then the

73 MHT shall be also relatively stable. This leads to the expectation that any large variations in heat
74 transport in the atmosphere and in the ocean should be equal in magnitude and opposite in sign,
75 nevertheless it does not rule out large changes in both ocean and atmosphere. This mechanism
76 since have been known as the Bjerknes compensation (BJC). Stone (1978) later showed that the
77 MHT is mainly determined by the solar constant, the axial tilt, the radius of the Earth and the
78 mean planetary albedo. Among these variables only the albedo is an internal parameter of the
79 atmosphere-ocean system, and it highly depends on temperature, as a defining factor on clouds,
80 ice and snow. Different modelling studies suggest that the BJC is valid in different paleo climate
81 states, for example during the glacial-interglacial period of the last 22 000 years (Yang et al.,
82 2015), or in the warm climate of the Middle Miocene (Krapp & Jungclaus, 2011). Also, it was
83 found in preindustrial and in historical simulations of CMIP5 models, where in the latter, the
84 climate is not in equilibrium and external forcings are present (Outten et al., 2018). Even in
85 extreme theoretical cases, such as the aqua-planet, BJC is shown to be valid (Smith et al., 2006).

86 Even though total MHT is stable in different climate states, the contributions from
87 various transport processes might change. Quantifying transport processes is a great tool to
88 identify the large scale features of different climate states and reveal any compensating
89 mechanism. In the climate system the heat is transported by the atmosphere and the ocean via
90 different mechanisms. In the tropical belt both the atmosphere and the ocean contribute to the
91 MHT equally, while at higher latitudes the atmospheric transport dominates (Masuda, 1988). At
92 lower latitudes the heat is transported mainly by the meridional overturning circulation (MOC),
93 which is represented by the Hadley cell in the atmosphere and by the wind-driven gyres in the
94 upper ~1000 m of the ocean (Held, 2001). Note that we do not evaluate separately the role of the
95 ocean's meridional overturning circulation so that MOC hereafter refers always to the
96 atmosphere in this study. At higher latitudes the heat is transported dominantly by the
97 atmosphere via different eddies, such as transient eddies (TE), comprised mainly of mid-latitude
98 cyclones, and stationary eddies (SE), which represent monsoons in the subtropics and planetary
99 waves in midlatitudes. The stationary planetary waves are connected to diverse topography and
100 land-sea thermal contrast (Wills et al., 2019).

101 Change in the different transport processes have been shown to be relevant in the present
102 climate, for example the changes in the poleward atmospheric and oceanic heat transport has
103 been shown to contribute to the polar amplification (Forster et al., 2021). In a warmer climate the
104 equator-to-pole gradient of atmospheric moisture is expected to increase due to the exponentially
105 increasing water holding capacity of warmer air. The larger moisture gradient enhances the
106 poleward latent heat transport. This process plays an important role in polar amplification. The
107 polar amplification on the other hand results in a weakening of the equator-to-pole temperature
108 gradient, which decreases the dry-static energy transport, what in turn partly compensates the
109 increase in latent heat transport (Forster et al., 2021). The latest IPCC report shows that in our
110 current warming climate large scale circulation patterns such as the Hadley cell or the monsoon
111 systems have been changing (Gulev et al., 2021). Regarding the Hadley circulation the studies
112 mostly indicate changes in the northern Hadley cell, in the form of widening and strengthening
113 of the circulation (Gulev et al., 2021). The North Hemisphere summer monsoon precipitation
114 shows an increase in the recent past (Gulev et al., 2021).

115 Investigating the large scale circulation trough, the transport processes in a past warm
116 period of Earth's climate system can help to better understand our future in a warmer climate.
117 Paleoclimate model simulations in contrast to future projections, have the advantage, that there

118 are proxy data available, which help to validate the model's response to any changes in the
 119 forcing. Proxies help reducing the uncertainties, when modelling a climate state in extreme
 120 circumstances. One of the best examples of past warm periods is the Early Eocene Climatic
 121 Optimum (EECO, ~56-48 Myr ago). It is the period of greatest sustained (> 1 Myr) warmth in
 122 the last 65 million years (Lunt et al., 2017), when CO_2 concentrations are estimated to have
 123 fallen between 1170 and 2490 ppm (Anagnostou et al., 2020) and the estimated global mean
 124 surface temperatures reached 27.0°C (23.2 to 29.7°C), approximately 10 to 16°C warmer than
 125 preindustrial climate (Inglis et al., 2020). It has been shown (Evans et al., 2018), that the
 126 meridional temperature gradient was much weaker in the Eocene's warm climate than at present,
 127 thus indicating strong polar amplification, which is a challenge to capture for most climate
 128 models. There has been a community effort in creating a framework for the intermodel
 129 comparison of Paleocene-Eocene simulations (Lunt et al., 2017) and also to coordinate the
 130 methodology of a proxy data compilation focusing on temperature and CO_2 concentrations from
 131 this period (Hollis et al., 2019). This coordinated effort is the Deep Time Model Intercomparison
 132 Project (DeepMIP), and the time intervals its focusing on are the latest Paleocene (pre-PETM),
 133 Paleocene–Eocene thermal maximum (PETM) and early Eocene climatic optimum (EECO). In
 134 DeepMIP, the atmospheric CO_2 concentrations and other boundary conditions, such as the
 135 paleogeography, orbital configurations, solar constant, vegetation, and lack of continental ice
 136 sheets were uniformly defined. Eight modelling groups participated in performing paleo
 137 simulations, with the agreed boundary conditions. Some models are closer to the proxy records
 138 than others, but all simulations indicate a warmer, than present day global mean temperature and
 139 a smaller, than present day meridional SST gradient. The Community Earth System Model
 140 (CESM), the Geophysical Fluid Dynamics Laboratory model (GFDL) and the Norwegian Earth
 141 System Model (NorESM) perform well in simulating the global mean temperature and
 142 meridional SST gradient at CO_2 levels, which are also in the proxy indicated range (Lunt et al.,
 143 2021). The closest agreement with proxy data is found for simulations at $6x$ times the
 144 preindustrial CO_2 concentration, which also aligns with the best-estimate of CO_2 signal from
 145 proxy data. In terms of simulating the lowest meridional temperature gradient the most
 146 successful simulation is from CESM (Lunt et al., 2021).

147 Several studies have investigated the EECO's heat transport with the intention to explain
 148 its low temperature gradient. Increased heat transport, especially increased ocean transport have
 149 been suggested to be the mechanism responsible for the low meridional gradient (Barron, 1987;
 150 Sloan et al., 1995). It has been estimated that 30% or even greater increase in poleward heat
 151 transport, than at present is needed to achieve the low temperature gradient of Eocene (Huber &
 152 Nof, 2006; Sloan et al., 1995). This is problematic, since we do not know any acceptable
 153 mechanism to realize such a large transport change, neither is it something any respective
 154 modelling study would have indicated. The Eocene meridional heat transport have been studied
 155 with coupled global circulation models (Heinemann et al., 2009; Huber & Sloan, 2001), and
 156 their results support the notion that not the increased oceanic heat transport is the responsible
 157 mechanism for the low meridional temperature gradient.

158 In our work we focus on the changes in the different atmospheric transport processes in
 159 the EECO, to better understand the atmospheric large scale circulation pattern of this warm
 160 climate. Moreover, we separate these changes depending on their underlying causes, namely if
 161 they are driven by the CO_2 increase or the non- CO_2 forcing of the paleo simulations. We
 162 compare the results of five different models from the DeepMIP ensemble. Dividing the changes
 163 into non- CO_2 and CO_2 forcing helps us to also assess the relevance of the results for future

164 climate scenarios, where the CO₂-driven processes become more relevant than any changes in
165 topography. Our study aims to answer the following questions:

- 166 • Can the DeepMIP model ensemble capture the characteristics of transport processes in
167 the preindustrial control (PI) simulations?
- 168 • What are the impacts of changes of non-CO₂ constraints (paleogeography, vegetation, no
169 continental ice sheet) on the different atmospheric transport processes?
- 170 • What are the impacts of CO₂ concentration increase on the atmospheric transport
171 processes in the 3x and 6x CO₂ EECO simulations?
- 172 • How does the overall change (non-CO₂ and CO₂ constraints) between the preindustrial
173 control and EECO simulations look like? Which physical transport processes are affected
174 the most?

175 The paper is structured as follows, in section 2 we briefly introduce the DeepMIP
176 experimental design, the selected models and reanalysis, and we explain the methods used in the
177 analysis. Then the results section shows the transport changes due to the non-CO₂ and the CO₂
178 constraints, first individually and then their combined effect on changes between past and present
179 climates. Then section 4 discusses the three large scale circulation patterns, which are affected by
180 either one or both of the CO₂ and non-CO₂ constraints. In section 5 we summarize our findings
181 and conclude.

182 **2 Data and Methods**

183 In this study we analyze climate model simulations from DeepMIP simulations. We
184 further include present day data from the ERA5 reanalysis to compare it with the respective
185 preindustrial simulations of the DeepMIP models.

186 **2.1 Experimental design**

187 The experimental design and the different models included in DeepMIP are described in
188 Lunt et al. (2017, 2021), here we only introduce them briefly. DeepMIP was conducted to offer a
189 consistent framework for climate model simulations of three warm periods in the latest
190 Paleocene and early Eocene (~ 55 to ~ 50 Ma), which are the Early Eocene Climatic Optimum
191 (EECO), the Paleocene–Eocene Thermal Maximum (PETM) and the period just before the
192 PETM (pre-PETM). These time periods of Earth’s climate are characterized by high atmospheric
193 CO₂ concentrations estimated to be between 800 and 3160 ppm (Anagnostou et al., 2020), which
194 means at least 3x higher concentrations than the preindustrial value (280 ppm) and almost 12x
195 more at its highest. The concentrations during the EECO, which is the longest of the three, is
196 estimated to fall between 1170 and 2490 ppm. This is around 4x and 9x the preindustrial
197 concentration level. Many DeepMIP groups performed multiple experiments at various CO₂
198 levels, for example at 1x, 3x, 6x or 9x the preindustrial CO₂ concentration, to capture this
199 uncertainty. Apart from the atmospheric CO₂ concentrations other boundary conditions, such as
200 the paleogeography, orbital configurations, solar constant, vegetation, continental ice sheets and
201 aerosols, are needed to be defined to set up a deep-time simulation. The paleogeography
202 represents the Ypresian stage of the Eocene, where the most notable differences to today’s
203 geography are the lack of the Himalaya, the lack of the enclosed Mediterranean basin, the proto-
204 Paratethys and the Siberian Sea and a narrower Atlantic basin. The digital reconstruction from

205 Herold et al. (2014) is used for the paleogeography, vegetation, and river routing. The orbital
 206 configuration is set to the modern values, because it represents a forcing close to the long-term
 207 average (Lunt et al., 2017). Both the solar constant, and non-CO₂ greenhouse gas concentrations
 208 are set to preindustrial values, to find a middle ground between the uncertainty on the increased
 209 radiative forcing associated with enhanced non-CO₂ greenhouse gases and the decrease in
 210 radiative forcing via a reduced solar constant (Lunt et al., 2017). One additional important initial
 211 constraint is that there are no continental ice sheets in the Eocene simulations. Initial condition
 212 for ocean temperature and salinity are given in Lunt et al. (2017) but each modeling group
 213 followed their individual approach based on their previous paleo simulations or experiences with
 214 model instabilities.

215 To accept a model simulation as the representation of the paleoclimate of EECO, PETM, or pre-
 216 PETM it needs to be in (or close to) equilibrium. To assure this, three criteria are defined, the
 217 first is regarding the length of the simulation (more than 1000 years) the second considers the
 218 energy balance of the models (less than 0.3 W m⁻² net radiation imbalance at the top of the
 219 atmosphere or similar imbalance to that of the preindustrial control), and last is the test if the
 220 ocean have reached its equilibrium state (sea-surface temperature trend is less than 0.1°C per
 221 century in the global mean). These latter two shall be based on the final 100 years of the
 222 simulation. Most of the simulations fulfill these conditions, and those which are not, only
 223 overstep the thresholds slightly, thus Lunt et al. (2021) concluded them to be sufficiently
 224 equilibrated.

225

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Table 1

List of models used in this study from the DeepMIP ensemble.

227

228

Model (short name)	Experiments	Length of simulations (years)	Atmospheric resolutions (lat x lon)
CESM (CESM1.2_CAM5)	piControl 1x,3x,6x	2000	1.9° x 2.5°
COSMOS (COSMOS-landveg_r2413)	piControl 1x,3x	9500	3.75° x 3.75°
GFDL (GFDL_CM2.1)	piControl 1x,3x,6x	6000	3° x 3.75°
HadCM3 (HadCM3B_M2.1aN)	piControl 1x,3x	7800	3.75° x 2.5°
MIROC (MIROC4m)	piControl 1x,3x	5000	2.79° x 2.81°

229

230

231 2.2 Models

232 In total 8 models participated in DeepMIP from which we selected 5, depending on the
 233 available experiment types. In our study we focus on the CO₂ and non-CO₂ effects, thus we

234 needed from all models a preindustrial control simulation, a 1x CO₂ Eocene simulation, and also
 235 at least one Eocene simulation with a higher CO₂ concentration. The three models, which we did
 236 not include in our study (INMCM, IPSL, NorESM), are left out because there was no available
 237 1x CO₂ simulation from them. We have chosen 3x and 6x CO₂ concentration simulations, where
 238 they were available (see Table 1). These concentrations, represent the pre-PETM and the
 239 EECO/PETM conditions respectively. All models are coupled ocean-atmosphere models. The
 240 selected simulations are summarized in Table 1. An overview of each model is listed below,
 241 while the more elaborate description of the simulations and models are found in Lunt et al.
 242 (2021) and in the corresponding papers.

243 CESM stands for the Community Earth System Model version 1.2, it consists of the
 244 Community Atmosphere Model 5.3 (CAM), the Community Land Model 4.0 (CLM), the Parallel
 245 Ocean Program 2 (POP), the Los Alamos sea ice model 4 (CICE), the River Transport Model
 246 (RTM), and a coupler connecting them (Hurrell et al., 2013). The atmospheric part of the
 247 coupled system has 30 hybrid sigma-pressure levels and a horizontal resolution of $1.9^\circ \times 2.5^\circ$
 248 (latitude \times longitude). The ocean and sea ice model use a nominal $1^\circ \times 1^\circ$ displaced pole
 249 Greenland grid with 60 vertical levels in the ocean. Some modifications were needed to make the
 250 Earth system model applicable for a paleoclimate simulation with a high CO₂ level applicable.
 251 These effected the radiation parametrization, and the marginal sea balancing scheme. The ocean
 252 was initialized from a previous PETM simulation (Kiehl & Shields, 2013) without any sea ice.
 253 The simulations have been integrated for 2000 model years, except the 1xCO₂ simulation,
 254 which has been integrated for 2600 model years.

255 COSMOS is developed at the Max Planck Institute for Meteorology and uses the
 256 atmospheric general circulation model ECHAM5 (Roeckner et al., 2003) and the Max-Planck-
 257 Institute for Meteorology Ocean Model (MPIOM) (Marsland et al., 2003) for the the ocean and
 258 sea ice components. The atmospheric part has 19 vertical hybrid sigma-pressure levels and a
 259 horizontal resolution approximately $3.75^\circ \times 3.75^\circ$. The ocean and sea ice dynamics are
 260 calculated on a bipolar curvilinear model grid with formal resolution of $3.0^\circ \times 1.8^\circ$ (longitude \times
 261 latitude) and 40 unequal vertical levels. COSMOS' performance in paleoclimate studies is
 262 described in Stepanek and Lohmann, (2012). The ocean was initialized with uniformly horizontal
 263 and vertical temperatures of 10 °C in the 3xCO₂ concentration simulation and then the
 264 simulations with 1 \times and 4 \times CO₂ concentrations were restarted from 3xCO₂ after 1000 years.
 265 The simulations were started with transient orbital configurations and were run for model year
 266 8000, then they were run with preindustrial orbital parameters for the last 1500 years.

267 GFDL means the Geophysical Fluid Dynamics Laboratory (GFDL) CM2.1 model
 268 (Delworth et al., 2006), with modifications to the late Eocene (Hutchinson et al., 2018, 2019).
 269 The CM2.1 consists of Atmosphere Model 2, Land Model 2, the Sea Ice Simulator 1. The ocean
 270 is calculated by the modular ocean model (MOM) version 5.1.0. The atmosphere has 24 vertical
 271 levels and a horizontal resolution of $3^\circ \times 3.75^\circ$. The ocean and sea ice components are
 272 calculated over 50 vertical levels with the horizontal resolution of $1^\circ \times 1.5^\circ$ (latitude \times
 273 longitude), and a tripolar grid is used as in Hutchinson et al. (2018). Due to the paleogeography
 274 some manual adjustments were made in the ocean grid. The ocean temperature was initiated
 275 from idealized conditions, similar to those outlined in Lunt et al. (2017). The simulations were
 276 run for a total of 6000 years. During the initial 2000 years, two adjustments were performed on
 277 the ocean temperature to accelerate the approach to equilibrium. This approach led to instabilities

278 at 6xCO₂ level, this simulation was instead initialized using a globally uniform temperature of
 279 19.32 °C and was run continuously for 6000 years.

280 HadCM3 is the abbreviation of the Hadley Centre Climate Model (Valdes et al., 2017).
 281 The atmosphere has 19 vertical levels and a horizontal resolution of 3.7° × 2.5° , while the ocean
 282 is calculated on a 1.25° × 1.25° grid over 20 vertical levels. A few changes were necessary to
 283 adapt the model to the deep-time simulations, such as a salinity flux correction, prognostic 1D
 284 ozone scheme instead the fixed vertical profile, and also the disabling of modern-day specific
 285 parametrizations, e.g., in the Mediterranean and the Hudson Bay. The ocean was initialized from
 286 the final state of Eocene model simulations using lower resolution in the ocean, HadCM3L. The
 287 HadCM3L simulations were initialized from a similar idealized temperature and salinity state as
 288 described in Lunt et al., (2017). HadCM3 simulations were started from the respective
 289 HadCM3L integrations after 4400 to 4900 years of spin up and run for a further 2950 years.

290 MIROC stands for Model for Interdisciplinary Research on Climate (Chan & Abe-Ouchi,
 291 2020). The land surface model is the Minimal Advanced Treatments of Surface Interaction and
 292 Runoff (MATSIRO) (Takata et al., 2003). The ocean component is the version 3.4 of the CCSR
 293 (Center for Climate System Research) Ocean Component Model (COCO) (Hasumi, 2000). The
 294 atmosphere has 20 vertical sigma levels and a horizontal resolution of approximately 2.79° x
 295 2.81° (latitude x longitude). The ocean has 44 levels and a horizontal resolution is set to 256 ×
 296 196 (longitude × latitude), with a higher resolution in the tropics. The atmosphere is initialized
 297 from a previous experiment without ice sheets and with a × 2 CO₂ concentration. The ocean is
 298 initialized based on previous MIROC paleoclimate experiments and on the recommendations
 299 from Lunt et al. (2017). For the experiments the model was run for 5000 model years.

300 2.3 Reanalysis

301 We use the atmospheric reanalysis ERA5 (Hans Hersbach et al., 2020) to evaluate the
 302 DeepMIP ensembles performance for the preindustrial control simulation. ERA5 is a
 303 comprehensive reanalysis from Copernicus Climate Change Service (C3S) produced by
 304 ECMWF and it is based on the Integrated Forecasting System (IFS) Cy41r2 which was
 305 operational in 2016. In our study we used monthly averaged data on pressure levels and on single
 306 levels (H. Hersbach et al., 2019b, 2019a) from the time period 1991-2020 on a horizontal
 307 resolution of approximately 0.25°x0.25°.

308 2.4 Partitioning the meridional heat transport

309 In our study we analyze the total meridional heat transport, and focus on its components,
 310 in the atmosphere. We follow the method described in Donohoe et al. (2020). First the MHT is
 311 partitioned between the ocean and the atmosphere

$$312 \quad MHT = OHT + AHT, \quad (1)$$

313 and the atmospheric transport is further partitioned into contributions from meridional
 314 overturning circulation (MOC), stationary eddies (SE) and transient eddies (TE)

$$315 \quad AHT = MOC + SE + TE. \quad (2)$$

316 Furthermore, all parts of Eq. (2) are divided into dry and moist energy transport (Eq. 3),
 317 where the moist part consists the transport of energy via latent heat and the dry part consists the
 318 transport of potential energy and sensible heat.

$$319 \quad AHT = AHT_{moist} + AHT_{dry}$$

$$320 \quad MOC = MOC_{moist} + MOC_{dry}$$

$$321 \quad SE = SE_{moist} + SE_{dry}$$

$$322 \quad TE = TE_{moist} + TE_{dry} \quad (3)$$

323 The total meridional heat transport at a latitude circle can be calculated with dynamic and
 324 energetic approaches, where in the energetic approach the MHT is balanced by the spatial
 325 integral of the net radiative deficit at the top of the atmosphere (TOA) and in the dynamic
 326 approach MHT is the vertically and zonally integrated net transport of energy. Here we use the
 327 energetic approach to calculate MHT (Eq.4)

$$328 \quad MHT(\Phi) = 2\pi a^2 \int_{-\frac{\pi}{2}}^{\Phi} \cos \Phi' [ASR(\Phi') - OLR(\Phi')] d\Phi', \quad (4)$$

329 where ϕ is the latitude circle, a is the radius of the Earth, ASR is the absorbed solar
 330 radiation, OLR is the outgoing longwave radiation. The boundary condition is that the transport
 331 has to be zero at the pole, because non-zero values have no physical meaning. To fulfill the
 332 boundary conditions at both poles, we need to balance the global budget. For this we assume that
 333 the imbalance is spatially uniform, thus the area-weighted global average energy imbalance is
 334 subtracted at all latitudes before calculating the integral in Eq. (4).

335 OHT can also be calculated with the energy approach from the surface heat fluxes (SHF)
 336 (Eq. 5)

$$337 \quad OHT(\Phi) = 2\pi a^2 \int_{-\frac{\pi}{2}}^{\Phi} \cos \Phi' [SHF(\Phi')] d\Phi', \quad (5)$$

338 where SHF is positive downward. The above equation is true with the assumption that the
 339 ocean is in equilibrium, so the heat storage is negligible. If the ocean is not in an equilibrium
 340 state, then Eq. (5) represents the implied OHT, which is the sum of OHT and the spatial integral
 341 of the tendency of ocean heat content. The DeepMIP simulations have been required to fulfill
 342 different criteria to prove that they reached the equilibrium state. The criteria considered the
 343 length of the simulations, the radiation imbalance at the TOA, and the ocean's equilibrium state.
 344 All of the included simulations satisfy at least two of the three criteria, with the only exception of
 345 CESM at 3xCO₂, which is still close to both missed criteria, thus all, here considered
 346 simulations, have been accepted to be sufficiently equilibrated (Lunt et al., 2021). We consider
 347 Eq.(5) defining OHT and that balancing the global radiative budget at the TOA does not largely
 348 affect largely the calculation of MHT.

349 After calculating MHT and OHT given Eq. (1) AHT is known as the residual. The direct
 350 dynamical calculation of AHT is formulated via the vertically and zonally integrated meridional

351 transport of moist static energy (MSE):

$$352 \quad MSE = c_p T + Lq + gZ$$

$$353 \quad AHT(\Phi) = \frac{2\pi a \cos \Phi}{g} \int_0^{P_s} [\bar{V}] [\overline{MSE}] + [V^* MSE^*] + [\overline{V'^* MSE'^*}] + [\overline{V'}] [\overline{MSE}]' dp,$$

354 (6)

355 where c_p is the specific heat capacity of air at constant pressure, T is temperature, L is the
 356 latent heat of vaporization of water, q is the specific humidity, g is the acceleration of gravity, Z
 357 is the geopotential height, P_s the surface pressure, V is the meridional velocity. The square
 358 brackets $[x]$ denote zonal averages, the overbars \bar{x} denote time averages (monthly means), x^*
 359 denote zonal anomalies, and x' means time anomalies. In Eq. (6) the first term defines the energy
 360 transported via MOC the second is SE, the third is TE, and the last one has been referred as the
 361 transient overturning circulation (TOC) (Marshall et al., 2014), which is two orders of
 362 magnitudes smaller than MOC at the tropics and the eddy terms at midlatitudes. Thus, we do not
 363 try to consider TOC on its own, but handle it together with TE. Note that the calculation of TE
 364 and TOC would require high temporal resolution data, which is not available for the DeepMIP
 365 simulations, thus we cannot calculate AHT only from Eq. (6), hence we use the residual method
 366 via Eq. (1). Nevertheless, the transport via MOC and SE is calculated from Eq. (6) with monthly
 367 mean data. The remaining atmospheric transport, which we refer to as TE, is again defined with
 368 a residual method. This TE calculation, has been shown to be successful in calculating the
 369 partitions from monthly mean data with good accuracy (Donohoe et al., 2020).

370 Regarding the moist and dry partitioning for AHT we define AHT_{moist} as the latent heat
 371 transport at a given latitude, which is the integral of evaporation (E) minus precipitation (P)
 372 multiplied by the latent heat of vaporization:

$$373 \quad AHT_{moist}(\Phi) = 2\pi a^2 \int_{\frac{\pi}{2}}^{\Phi} \cos \Phi' \{L[E(\Phi') - P(\Phi')]\} d\Phi' . \quad (7)$$

374 The dry contribution to AHT is then calculated by subtracting the moist part from the
 375 total (Eq. 3). The moist and the dry parts of MOC and SE are calculated via Eq.(6), but MSE has
 376 been split into dry, the sensible heat and potential energy ($c_p T + gZ$) and moist part, the latent heat
 377 (qL) (Donohoe et al., 2020). The moist and dry contributors to TE can be calculated via the
 378 residual method, with the use of Eq. (3) and:

$$379 \quad AHT_{moist} = MOC_{moist} + SE_{moist} + TE_{moist}.$$

380 2.5 Meridional Streamfunction

381 There are several metrics to quantitatively analyze the Hadley cell, its edge and its
 382 circulation strength. We choose to use the average meridional streamfunction Ψ (Xian et al.,
 383 2021) to quantify its intensity, which is defined as :

$$384 \quad \Psi(p, \Phi) = \frac{2\pi a \cos \Phi}{g} \int_{P_s}^0 [v] dp .$$

385 A stronger streamfunction means a stronger Hadley cell circulation.

386 2.6 Monsoon Area

387 The monsoon climate is characterized by seasonal reversal of prevailing surface winds
388 which results in a rainy summer and dry winter. To assess the geological area where monsoon
389 was probably present in the EECO, we use a simple monsoon definition, which is also used in
390 the 6th Assessment Report of IPCC (IPCC, 2021). The global monsoon is defined as the area,
391 where the local annual range of precipitation exceeds 2.5 mm per day (Kitoh et al., 2013). The
392 annual range is defined by the local summer- minus-winter precipitation, i.e. MJJAS minus
393 NDJFM in the Northern Hemisphere and NDJFM minus MJJAS in the Southern Hemisphere.
394 The 2.5 mm per day threshold is defined based on current conditions, which might not hold so
395 well in a warmer world, nevertheless we accept it for our rough estimations. A more detailed
396 analysis of the monsoon systems is to follow this study, where not just such simple indices will
397 be used.

398 **3 Results**

399 3.1 Preindustrial control simulation

400 We start our analysis by evaluating the selected models' representation of present day
401 climate, more precisely the climate of the preindustrial period. To assess the models' results we
402 compare them to transport values calculated from the ERA5 reanalysis from 1991-2020. Note
403 that this comparison is not entirely fair since the models and the reanalysis represent two
404 different climate periods approximately 150 years apart, and in this one and a half century the
405 climate constraints, especially the CO₂ concentration have changed. Nevertheless, both models
406 and reanalysis represent a climate with current topography, continental ice sheets and relatively
407 similar meridional temperature gradient, when compared to our knowledge of EECO's climate.
408 Thus, we accept these discrepancies and focus on the structure of the transport processes, and not
409 on the quantitative amounts.

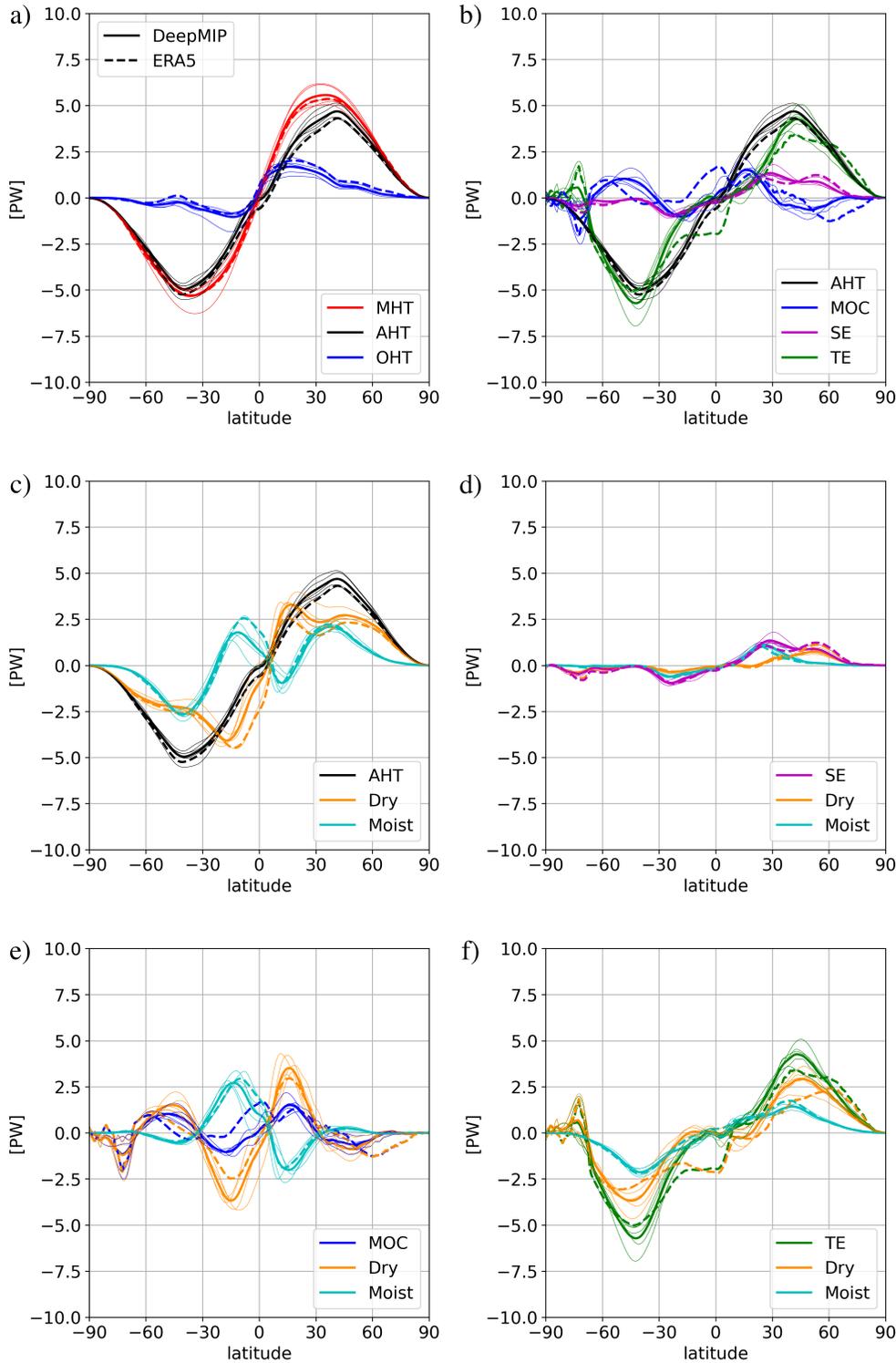
410 The meridional heat transport and its partitions calculated from the DeepMIP models and
411 the ERA5 reanalysis are shown in Figure 1. Overall, there is good agreement between the mean
412 of the model ensemble and the reanalysis. The values and distributions of the MHT and its
413 components fit well also to previous studies which are based on observations and reanalysis
414 (Donohoe et al., 2020; Masuda, 1988).

415 MHT reaches maximum of above 5 PW at around 40° both South and North. The ocean
416 transports more heat than the atmosphere in a narrow tropical belt (0-10 °N), outside of which
417 the atmospheric heat transport dominates (Figure 1a). The reanalysis is within the spread of the
418 DeepMIP ensemble. The ensemble mean fits the reanalysis better over the southern hemisphere
419 than over the northern, where the models transport more via the atmosphere and less via the
420 ocean. These two compensate each other resulting in a good fit for the total MHT.

421 The atmospheric transport partitioned into latent heat and dry static energy transport
422 (Figure 1c) shows that the two have similar order of magnitude, but the dry static energy
423 transport is at all latitudes poleward, while the latent heat transport is equatorward in the tropics
424 and poleward at the rest of the latitudes.

425 Partitioning of the atmospheric heat transport into meridional overturning circulation,

426 stationary, and transient eddies shows that transport is (a) mainly in transient eddies in the
427 midlatitudes, i.e. in extratropical cyclones, and (b) dominantly by meridional overturning
428 circulation, i.e., by the Hadley cells, in the tropics (Figure 1b). The tropical belt is where some
429 differences between the models and the reanalysis arises, namely over the Southern Hemisphere
430 the reanalysis shows more northward transport via meridional overturning circulation, MOC, and
431 more southward transport via transient eddies, TE. Since these two mechanisms act in the
432 different direction, the AHT does not show large differences between the reanalysis and the
433 ensemble mean. The separation of the MOC into its dry and moist parts (Figure 1e) reflects the
434 poleward transport of potential and sensible heat in the upper part of the Hadley cell and the
435 equatorward latent heat transport in the lower part of the Hadley cell. The transport of the two
436 parts do not balance out each other resulting in a net poleward energy transport. Stationary
437 eddies show importance in the poleward latent heat transport around 30° in both hemispheres,
438 representing the monsoon systems (Figure 1d). Stationary eddies also transport dry static energy
439 to the northern midlatitudes, mainly during winter, representing planetary waves (Masuda,
440 1988). Extratropical cyclones transport heat in both, the moist and dry form (Figure 1b and f).



441

442 **Figure 1.** Annual meridional heat transport and its different parts in the preindustrial control
 443 simulations of the 5 DeepMIP ensemble members and the ERA5 reanalysis. Bold lines
 444 representing the ensemble mean thin lines representing each model and dashed lines are
 445 calculated from ERA5 (1991-2020) reanalysis. The partitions of MHT are (a) Meridional Heat
 446 Transport, divided between atmospheric (AHT) and oceanic (OHT) part, (b) atmospheric

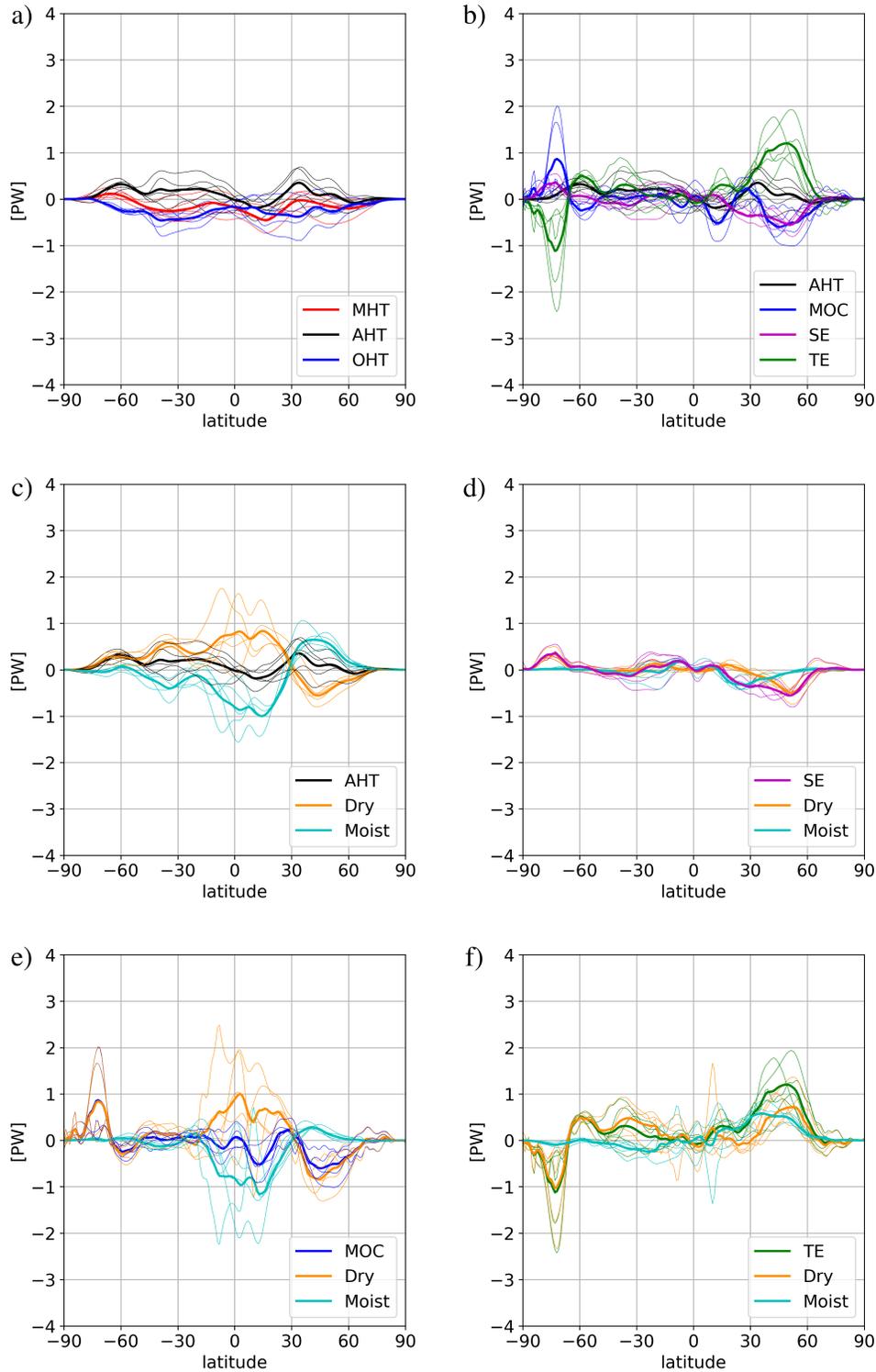
447 transport divided into Meridional Overturning Circulation (MOC), Stationary Eddies (SE) and
448 Transient Eddies (TE), (c) atmospheric transport divided to dry and moist parts, (d) atmospheric
449 transport via SE and its dry and moist parts, (e) atmospheric transport via MOC and its dry and
450 moist parts, and (f) atmospheric transport via TE and its dry and moist parts.

451 3.2 Effect of the Non-CO₂ forcing on the Eocene MHT

452 The Eocene 1xCO₂ simulations, i.e., with non-CO₂ forcing because of changes in
453 paleogeography and vegetation or the lack of continental ice sheets, are 3-5 °C warmer than the
454 preindustrial simulations (Lunt et al., 2021). Figure 2 shows the MHT differences between the
455 Eocene 1xCO₂ and the preindustrial control simulation. There is more southward MHT in the
456 1xCO₂ simulation than in the preindustrial one, which is mainly due to an increase in the oceanic
457 southward heat transport (Figure 2a). This results in smaller poleward MHT in the Northern
458 Hemisphere but larger in the Southern Hemisphere during the Eocene. In other words, due to the
459 Eocene boundary conditions the atmospheric poleward transport increases and the ocean
460 transport decreases in the Northern Hemisphere and vice versa in the Southern Hemisphere.

461 The change in atmospheric dry and moist transport is also asymmetric. North from the
462 Tropic of Cancer (30°N) more latent heat is transported northward and south from it more latent
463 heat is transported southward in the Eocene than in the preindustrial climate. The dry static heat
464 transport changes show the opposite sign. When taking into consideration the direction of
465 transport (see Figure 1c), this means that over the Northern Hemisphere there is more moisture
466 transport equatorward in the tropics and more poleward transport at midlatitudes, while the dry
467 static energy transport compensates these changes, with more poleward transport at the tropics
468 and less at higher latitudes (Figure 2c). On the other hand, over the Southern Hemisphere the
469 tropical equatorward latent heat transport decreases and the extratropical poleward latent heat
470 transport increases. The poleward dry static energy transport decreases at all southern latitudes.
471 These together lead to a net increase (decrease) in atmospheric poleward transport in the
472 Northern (Southern) Hemisphere, which as mentioned above is overcompensated by the ocean.
473 An increase in poleward latent heat transport also has an important role in polar amplification.

474 The most prominent atmospheric change is the increased heat transport by transient
475 eddies between 30°N-60°N and 60°S-90°S (Figure 2b). These represent heat transport by
476 extratropical cyclones, thus cyclones are more frequent and/or more intense in the Eocene
477 simulations. To quantify the number of cyclones and their features, more frequent than monthly
478 model output would be needed. The increased transient eddies transport is mostly compensated
479 by stationary eddies and meridional overturning circulation transports (Figure 2b). There is also a
480 decrease in the northern hemisphere monsoonal transport (Figure 2d). The Hadley circulation
481 also shows an asymmetric shift, with more energy being overturned in the northern cell, and less
482 in the southern one, while the net poleward MOC energy transport stays close to the control
483 simulation (Figure 2e).



484

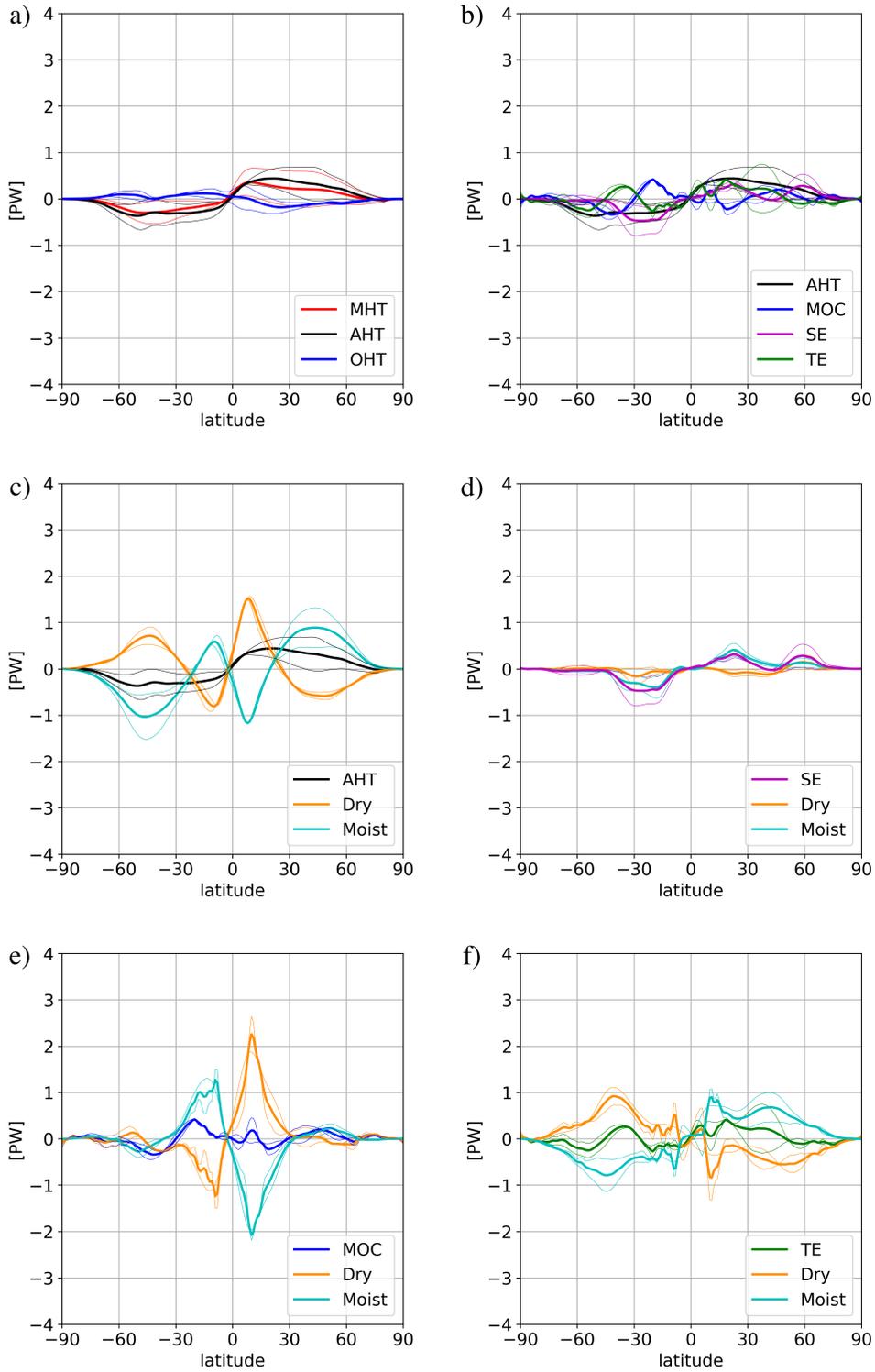
485 **Figure 2.** Meridional Heat Transport and its parts, same as Figure 1, but showing the differences
 486 between the 1xCO₂ Eocene and the preindustrial control simulations.

487

488 3.3 Effect of the CO₂ forcing on the Eocene MHT

489 The effect of higher CO₂ concentrations on early Eocene transport processes is studied by
490 quantifying the changes between the 3xCO₂ and 6xCO₂ simulations relative to the 1xCO₂
491 simulation (Figure 3 and 4). We compare the 6xCO₂ simulation also to the 1xCO₂ one to display
492 the intensification of the signal due to CO₂ rise. Our chosen DeepMIP ensemble has five
493 models available with 3xCO₂ simulations and only two with 6xCO₂ simulations (CESM &
494 GFDL, see Tab. 1). The two latter simulations are shown to be the most successful in
495 representing the global mean surface temperature, global mean SST and global meridional SST
496 gradient when compared to proxy data (Lunt et al., 2021) and thus we trust this small ensemble
497 is still representative.

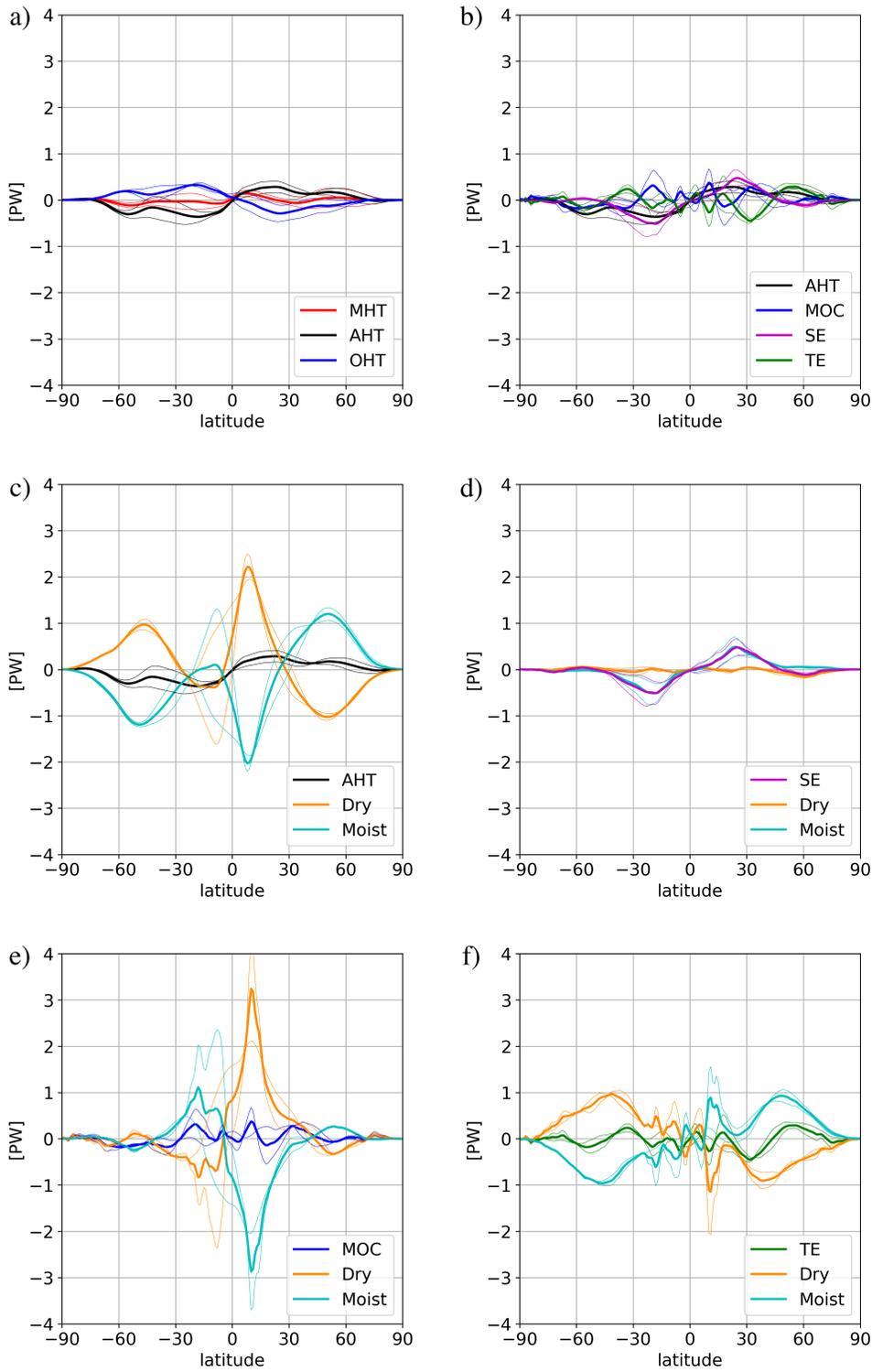
498 When comparing the changes relative to the equator, we see mostly symmetric changes,
499 thus the effect of CO₂ rise is global and influences both hemispheres in a similar way. For the
500 3xCO₂ simulation we find an increase in poleward MHT compared to the 1xCO₂ simulation,
501 which mainly results from the atmosphere (Figure 3a), while at the 6xCO₂ simulation the MHT
502 is fluctuating around zero (Figure 4a). In the latter case anomalies in the atmospheric and ocean
503 heat transport counteract one another, i.e., the Bjerknes compensation is present. The differences
504 between the two figures, and hence CO₂ concentrations, can also arise from the different set of
505 models used for the ensemble mean. Especially, since GFDL presents a non-linear behavior, and
506 from 4xCO₂ to 6xCO₂ concentration a slight decrease in the full MHT is found over the Northern
507 Hemisphere. Its magnitude is smaller than the CESM signal, thus the ensemble mean still
508 indicates a positive poleward transport change. Nevertheless, both at 3xCO₂ and 6xCO₂
509 concentration, the poleward atmospheric heat transport increases and ocean's heat transport
510 decreases or stays close to the 1xCO₂ value in the ensemble mean. Polar amplification is also
511 represented via the increased latent heat transport from the subtropics towards the poles,
512 compensated by the dry static heat transport (Figure 3c and 4c). Regarding the different physical
513 processes in the atmosphere, we see similar changes in both 3xCO₂ and 6xCO₂ simulations.
514 (Figure 3b and 4b). At the tropics the change in the net meridional overturning circulation
515 transport is close to zero, but the dry static and latent heat energy transport of the Hadley cell
516 increases more so at the northern cell than at the south (Figure 3e and 4e). The subtropics, are
517 mostly defined by the increased poleward transport of moist stationary eddies, namely the
518 transport of the monsoon systems, more in the 6xCO₂ than in the 3xCO₂ simulation (Figure 3d
519 and 4d). At midlatitudes the poleward transport of transient eddies increases slightly, especially
520 in the 6xCO₂ simulations (Figure 4f), but the magnitude of change is smaller than, what the non-
521 CO₂ constraints caused in the 1xCO₂ simulations (see Figure 2f).



522

523

524 **Figure 3.** Meridional Heat Transport and its parts, same as Figure 2, but showing the differences
 525 between the 3xCO₂ and the 1xCO₂ paleo simulation.



528 **Figure 4.** Meridional Heat Transport and its parts, same as Figure 3, but showing the
 529 differences between the 6xCO₂ and the 1xCO₂ paleo simulation. The ensemble here consists only
 530 two models: CESM, GFDL.

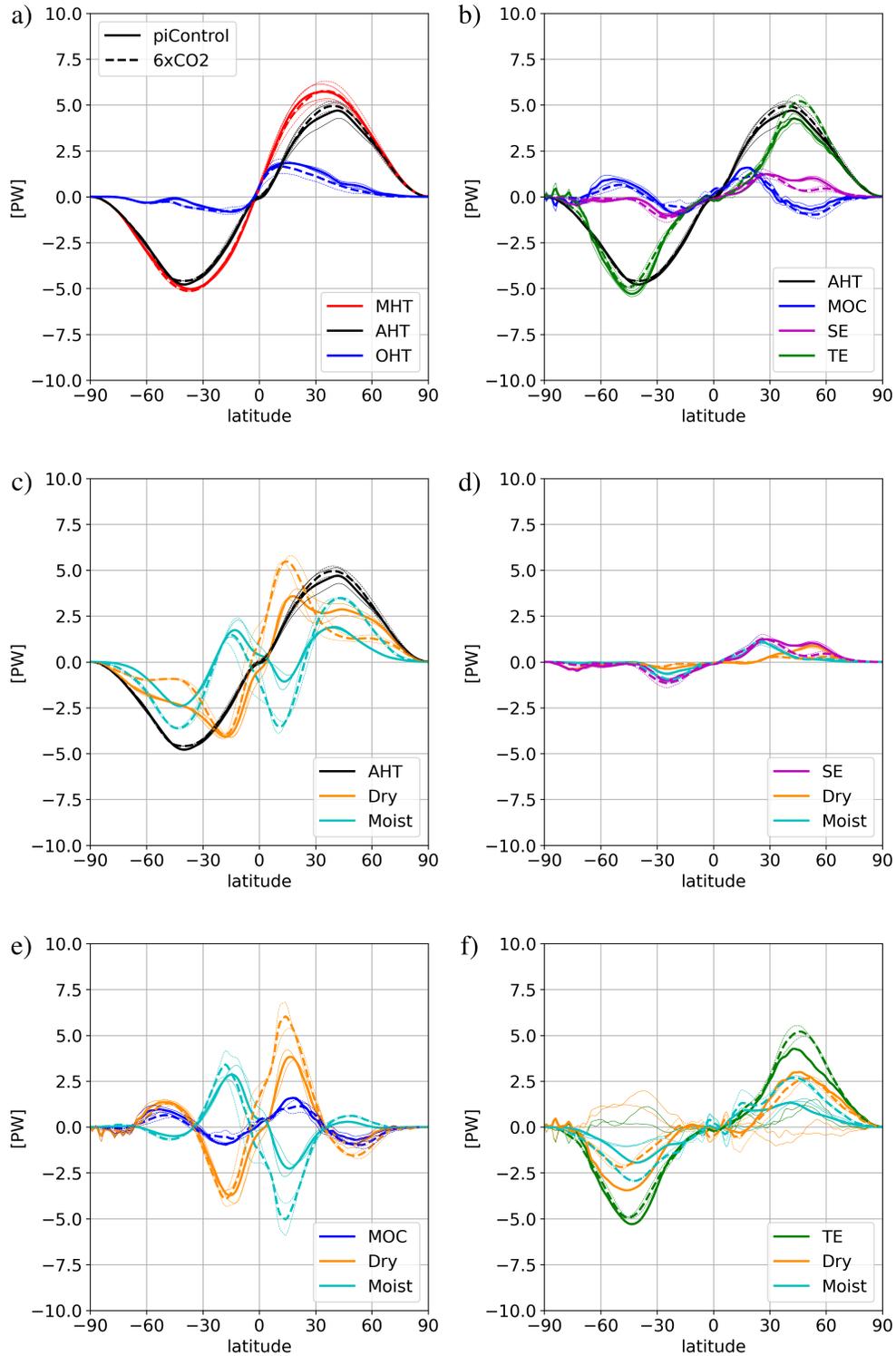
531

532 3.4 Effect of the total forcing on the Eocene MHT

533 Finally, we investigate the overall heat transport changes between the simulated
534 preindustrial and most-likely Eocene climate states. This is important as it is the relevant change
535 when comparing Eocene proxy data to present day conditions, and because the individual
536 changes due to the CO₂ and the non-CO₂ constraints are potentially nonlinear and can counteract
537 each other.

538 In total meridional heat transport, we see more changes compared to the present day
539 climate in the Northern Hemisphere, where the atmosphere transports more energy while the
540 ocean compensates this with less transport (Figure 5a). There is an increase of latent heat
541 transport toward the polar regions at both hemispheres in the EECO climate, which is
542 compensated by the decrease in dry static energy transport at the midlatitudes (Figure 5c). In the
543 tropics the net transport via meridional overturning circulation stays close to the preindustrial
544 values, but the dry (poleward) and moist (equatorward) energy transport increases in the Hadley
545 cell, especially in the northern cell (Figure 5e). In the southern subtropics the EECO simulations
546 show slightly more latent heat transport via stationary eddies (Figure 5d). This means that the
547 monsoon systems in the Southern Hemisphere transported more energy during the EECO.
548 Nevertheless, it has been shown in the previous two subsections that the monsoon in the
549 Northern Hemisphere also changed, but with opposite signs due to the respective CO₂ and non-
550 CO₂ forcings. Thus, the overall monsoonal transport changes in the Northern Hemisphere are
551 small. At the midlatitudes, especially in the Northern Hemisphere more energy is transported via
552 transient eddies (cyclones) during the Eocene, which is again compensated by less transport via
553 stationary eddies (Figure 5b,d,and e). In the Southern Hemisphere we find slightly less transient
554 eddy transport in the EECO than in the preindustrial simulations.

555



556

557

558 **Figure 5.** Meridional Heat Transport and its parts, same as Figure 1, but showing the
 559 preindustrial control (solid lines) and 6xCO2 (dashed lines) simulation results. The included
 560 models are: CESM, GFDL.

561 **4 Discussion**

562 The analysis of the different transport processes identified those large scale circulation
 563 patterns, which are either effected by the changes in the paleo set up or by the changes in the
 564 CO₂ concentration. These, heading from the tropics to the pole, are the Hadley cell, the monsoon
 565 and the mid-latitude cyclones. In this section we discuss the changes in these large scale patterns
 566 in more detail.

567 **4.1 Hadley Cell**

568 We found that due to both the CO₂ and non-CO₂ constraints more dry static energy and
 569 latent heat is turned around by the Hadley cells. Since all these simulations have a generally
 570 warmer climate than the control simulation it means that part of the surplus of energy compared
 571 to the preindustrial simulation, is coming from the higher water holding capacity of the warmer
 572 atmosphere. Both the CO₂ and non-CO₂ constraints cause a larger change in the Northern Hadley
 573 cell. When using the streamfunction for investigation of the Hadley circulation in the high CO₂
 574 simulations, all models show a weakening intensity of the Southern cell and a shift of the
 575 Northern cell towards south. The intensity of the Northern cell is strengthening in HadCM3,
 576 MIROC, and COSMOS, while CESM and GFDL show a slight weakening (see Table 2). Figure
 577 6 shows this exemplarily for the simulation GFDL. The southern cell is decreasing in intensity
 578 with the CO₂ rise, while the northern cell is expanding, mostly southward, and slightly increases
 579 in intensity. The southward expansion of the northern cell can be partially explained by the
 580 paleogeography. In the Eocene less continental land areas were located in the northern tropical
 581 belt. This could lead to a relative southward shift of the Intertropical Convergence Zone (ITCZ),
 582 given that during the summer half year the ITCZ travels poleward mostly over the continental
 583 areas. Nevertheless, the reason for the hemispheric asymmetry in overturned energy, intensity
 584 and position between the northern and southern Hadley cell due to CO₂ constraints is not entirely
 585 clear. These findings are in line with what we see in the modern day changing, warming climate.
 586 The 6th IPCC report states that a widening of the Hadley cells has been observed in the last
 587 decades, together with a strengthening of their circulation, especially in the Northern Hemisphere
 588 (Gulev et al., 2021).

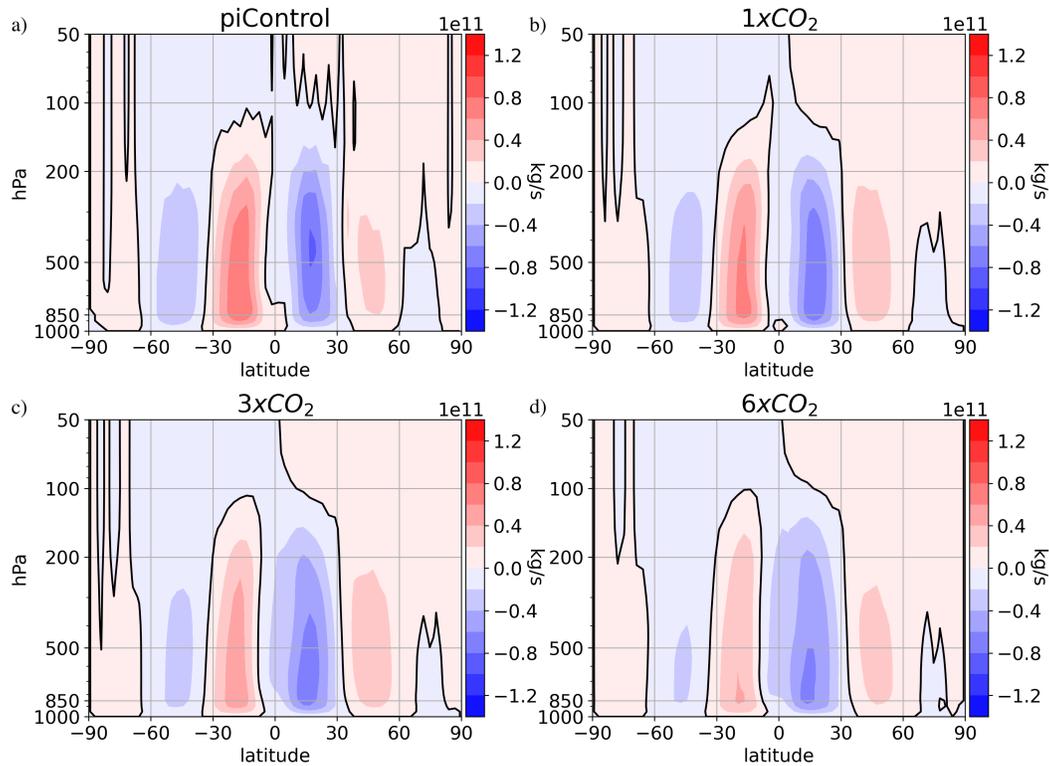
589

590 **Table 2**

591 *Streamfunction maximum and minimum values [kg/s] in the different models and*
 592 *simulations, indicating the intensity of each Hadley cell.*

	CESM		GFDL		HadCM3		COSMOS		MIROC	
	South	North								
piControl	1,0E+11	-8,8E+10	7,8E+10	-8,2E+10	9,2E+10	-9,6E+10	9,1E+10	-7,3E+10	1,1E+11	-7,3E+10
1xCO ₂	6,1E+10	-7,4E+10	6,7E+10	-7,6E+10	9,0E+10	-1,4E+11	8,7E+10	-9,2E+10	9,2E+10	-8,8E+10
3xCO ₂	5,5E+10	-6,7E+10	5,1E+10	-6,8E+10	7,9E+10	-1,3E+11	8,1E+10	-8,6E+10	8,4E+10	-9,9E+10
6xCO ₂	5,4E+10	-6,6E+10	4,2E+10	-6,5E+10						

593



594

595 **Figure 6.** Cross section of the annual mean meridional streamfunction in the (a)
596 preindustrial control, (b) 1xCO₂, (c) 3xCO₂, and (d) 6xCO₂ simulations of the GFDL model.

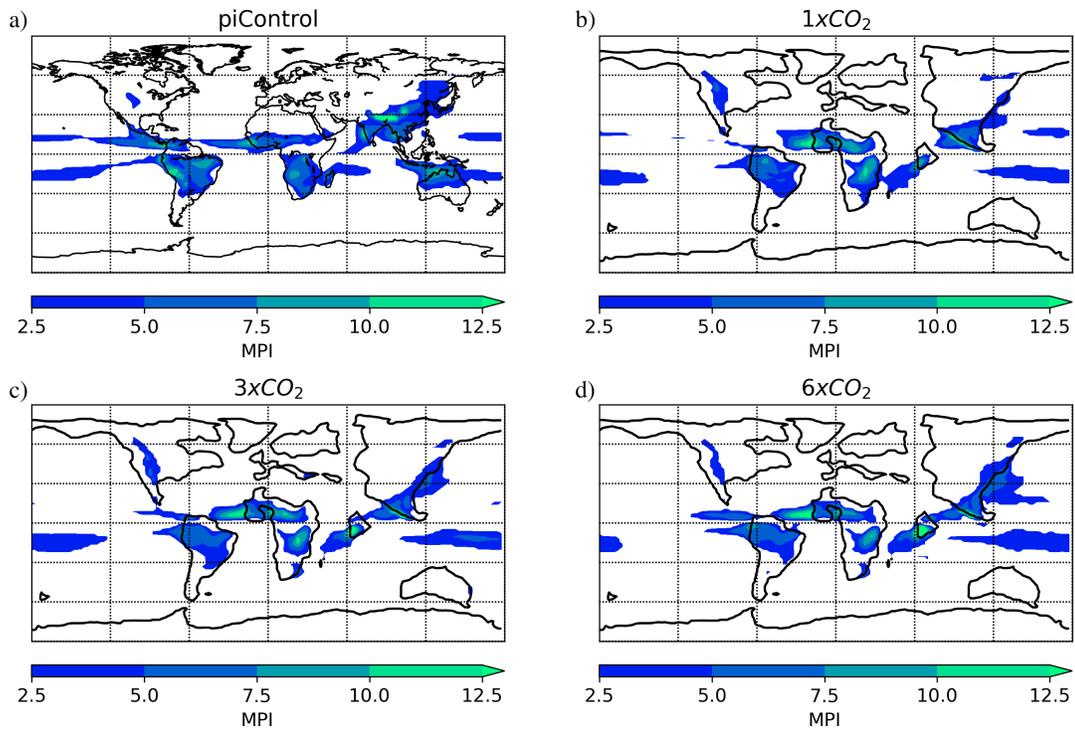
597 4.2 Monsoon

598 In the subtropics the transport via moist stationary eddies represents the heat transport by
599 monsoon systems. The analysis of the non-CO₂ effects showed a decrease in their transport in the
600 Northern Hemisphere and the CO₂ effects analysis showed an increase in transport with the CO₂
601 rise. Thus, we investigated the monsoon area to better evaluate these changes. The monsoon
602 area is defined by a precipitation index in all simulations.

603 In Figure 7 the monsoon areas are plotted from the CESM preindustrial 1x, 3x and 6xCO₂
604 concentration simulation. Paleogeography plays an important role in defining monsoon areas as
605 is shown by the smaller monsoon area in the early Eocene set-up than in the preindustrial. From
606 the preindustrial to the 1xCO₂ simulations, the percentage of monsoon areas decreases in all
607 models (see Table 3). The mean of the ensemble shows that in the preindustrial simulations the
608 monsoon covers 18.7% of the globe, while in the 1xCO₂ this area is reduced to 14.8%. This
609 correlates well with the slight decrease in the transport of moist stationary eddies in the
610 subtropics (Fig. 2d). From the 1x to the 3x and to the 6xCO₂ simulations (for CESM), the
611 monsoon area increases with higher CO₂ values (Figure 7). This also correlates well with the
612 results seen in the transport figures, where there is an increase in moist stationary eddy transport
613 in the subtropics (Figure 3d and 4d). The only exception from the increase of monsoon area with
614 CO₂ rise, is GFDL with a decrease in monsoon area from the 3x to the 6xCO₂ simulation.

615 When comparing the preindustrial climate to the EECO, namely the preindustrial

616 simulations to the 6xCO₂ simulations for CESM and GFDL simulations, we see that the effect of
 617 the non-CO₂ constraints (smaller monsoon area in the Eocene) and the effect of the CO₂ forcing
 618 (higher water holding capacity of the warmer atmosphere) compensate each other in terms of the
 619 energy being transported. This agrees well with the findings of Licht et al. (2014), who
 620 investigated proxy data from Asia in the late Eocene (gastropod shells and mammal teeth from
 621 Myanmar, and aeolian dust deposition in northwest China) and found monsoon like patterns in
 622 rainfall and wind. They also concluded that the enhanced greenhouse gas concentrations
 623 compensated the negative effect of lower Tibetan relief on precipitation, which reduces the East
 624 Asian monsoon strength (Tang et al., 2013). In summary, when comparing present and EECO
 625 monsoon systems from the energetic point of view, there is no large difference (Figure 5d),
 626 nevertheless the reason for this is due to compensating mechanisms. Note that this does not
 627 indicate that there is no significant change in monsoon precipitation intensity.



628
 629 **Figure 7.** Monsoon area defined by the Monsoon Precipitation Index (MPI) with the unit of
 630 mm/day : a) preindustrial control b) 1xCO₂ c) 3xCO₂ and d) 6xCO₂ simulation of the CESM
 631 model.

632
 633 **Table 3**
 634 *Global monsoon area in percentages in the different models and simulations.*

	CESM	GFDL	HadCM3	MIROC	COSMOS	ENS
piControl	18.83	21.87	17.97	16.36	18.45	18.70
1xCO₂	14.61	19.17	13.59	10.47	16.43	14.85

635

3xCO₂	14.91	19.95	14.26	12.75	22.43	16.86
6xCO₂	15.42	18.04	-	-	-	16.73

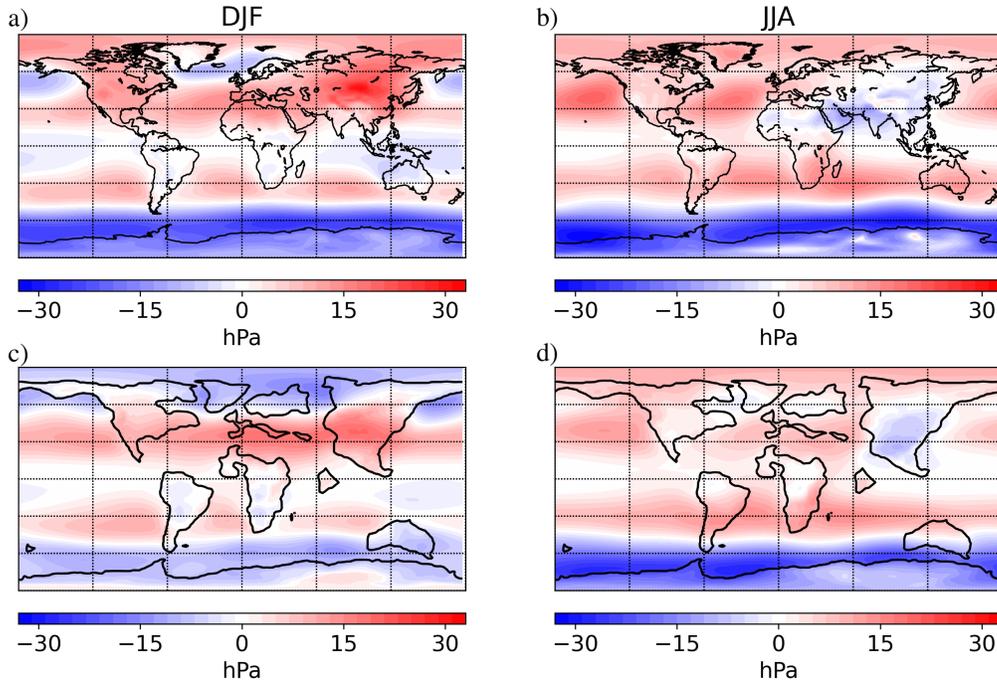
636

637 4.3 Midlatitude cyclones

638 We found that in the paleo set up even without the CO₂ increase, more energy is being
 639 transported via midlatitude cyclones especially in the Northern Hemisphere (Figure 2b). This can
 640 be explained by an increase in semi-permanent low pressure systems, in other name centers of
 641 action, in the Eocene simulations. In Figure 8 the sea level pressure anomalies are representing
 642 these centers of action. In the Northern Hemisphere in the preindustrial simulation one can
 643 identify two main low pressure systems during winter (Figure 8a), the Icelandic and the Aleutian
 644 lows, while in the Eocene even four low pressure centers can be identified. We call these the
 645 Icelandic, the Aleutian, the Gulf of Alaska and the Eurasian Low (Figure 8c).

646 The development of these semi-permanent pressure features is connected to the thermal
 647 contrast between the ocean and the continent during winter, due to the different heat capacity of
 648 land and sea. This also explains why the semi-permanent pressure systems develop differently in
 649 the paleo set up. In the Eocene world there was a wider Pacific basin and a narrower Atlantic
 650 basin together with the existence of the Turgai Sea or West Siberian Sea, an epicontinental sea
 651 separating Europe from Asia. The existence of the West Siberian Sea, which is located in the
 652 northern midlatitudes, lead, in the models, to the development of an Eurasian Low pressure
 653 system that is not in present in the modern times. The wider Pacific basin and the presence of the
 654 Bering land bridge lead, in the models, to a split in the Aleutian Low, so in the Eocene both an
 655 Aleutian Low and a Gulf of Alaska Low are present in the simulations. In the Southern
 656 Hemisphere the position of the Antarctic continent did not change much and, thus, the pressure
 657 systems in the models developed similarly as in the present climate (Figure 8 b and d).

658 We hypothesize that the increase in the energy transport via transient eddies in the
 659 paleosimulations are due to the increase in cyclonic activity due to more semi-permanent low
 660 pressure systems over the northern midlatitudes. This can mean more and/or deeper cyclones
 661 than in the present climate. To quantitatively assess this, the model output in our study with the
 662 monthly temporal resolution is insufficient.



663

664 **Figure 8.** Winter (left) and summer (right) sea-level pressure anomaly in the a) b)
 665 preindustrial control and c) d) 1xCO₂ paleosimulation of the CESM model.

666

5 Conclusions

667

668 In this study we calculated and analyzed the meridional heat transport and its partition in
 669 the atmosphere in climate model simulations of the preindustrial and the Early Eocene Climatic
 670 Optimum (EECO). We used simulations from five climate models (CESM, COSMOS, GFDL,
 671 HadCM3, and MIROC) provided by the DeepMIP community. The transport values are
 672 calculated from monthly mean data, and we distinguish between the different physical
 673 mechanisms, which transport energy in the atmosphere. The impacts of the non-CO₂ related
 674 conditions (paleogeography, vegetation, no continental ice sheets) and the CO₂ concentration
 675 forcing on the transport processes are calculated first separately, and then in combination, to
 676 allow a full comparison of heat transport in the preindustrial and early Eocene. The transport
 677 processes via the Hadley cell, the monsoon systems and the midlatitude cyclones are analyzed in
 678 more detail as these large-scale circulation patterns are identified as being different in the EECO
 compared to present day.

679

680 Our first research question investigates the DeepMIP models' skill in capturing the
 681 characteristics of transport processes in the preindustrial simulations (section 4.1). Overall, the
 682 DeepMIP ensemble mean and the reanalysis transport values show good agreement. Only the
 683 Southern Hemispheric tropical MOC transport shows marked differences, but due to two
 684 compensating mechanisms the overall atmospheric heat transport is still similar between the
 reanalysis and the multi-model mean.

685

686 The impacts of non-CO₂ constraints (paleogeography, vegetation, no continental ice
 687 sheet) on the different transport processes show hemispheric asymmetry. In the 1xCO₂ Eocene
 simulations the total poleward MHT is smaller in the Northern Hemisphere but larger in the

688 Southern Hemisphere. This is mainly due to the same change in oceanic transport. This shift in
 689 oceanic transport towards the South Pole is in line with the strong Southern Hemisphere-driven
 690 oceanic overturning circulation of the Eocene, found in the DeepMIP models (Zhang et al.
 691 (2022)). Considering the different physical processes most notable is the extra heat transport by
 692 transient eddies in the midlatitudes, which is compensated by a loss in energy transport via
 693 stationary eddies. The increase in cyclonic heat transport is explained by changes in
 694 paleogeography. The existence of an extra epicontinental sea at the north midlatitudes, with
 695 possible high land-sea thermal contrast, results in a semi-permanent pressure system, which
 696 impacts the transient eddies at the midlatitudes. Also, the wide Pacific basin results in a split in
 697 the Aleutian Low. Thus, in the end during the Eocene winter there were probably 4 semi-
 698 permanent low pressure systems present at the northern midlatitudes as opposed to the only two
 699 in present day climate. The increase in semi-permanent lows likely results in an increase in
 700 cyclone numbers as well, but to quantify this, one needs high temporal resolution output from the
 701 models. A decrease in North Hemispheric monsoon latent heat transport, is also linked to the
 702 change in topography, since the location of land areas highly determines the monsoon area.

703 Transport changes, which are solely due to CO₂ increases in the 3x and 6x CO₂ EECO
 704 simulations, are especially relevant to better understand and predict changes under future climate
 705 change. We find that with increasing CO₂ concentrations, the atmosphere transports more heat
 706 poleward, while the ocean transports the same amount or less. The results also show that the
 707 Hadley cell overturns more heat on the northern side of the Equator more than on the southern
 708 side. Also, monsoon systems transport more latent heat from the subtropics to the higher
 709 latitudes. This indicates that with CO₂ rise the hydrological cycle intensifies. This agrees well
 710 with what have been found in our current changing climate. At the end of the 20th century the
 711 Hadley circulation is shown to be strengthening particularly in the Northern Hemisphere, while
 712 the global monsoon precipitation shows a positive trend in the Northern Hemisphere (Gulev et
 713 al., 2021).

714 Our third research question considers the total change between the preindustrial control
 715 and EECO simulations, and asks which physical processes are affected the most. In our results
 716 we see more changes in the Northern Hemisphere, where the atmosphere transports more heat
 717 poleward while the ocean compensates this with less poleward heat transport. We found an
 718 increase in latent heat transport towards the polar regions in both hemispheres midlatitudes in the
 719 EECO climate, and a decrease in poleward dry static energy transport, same results as
 720 Heinemann et al. (2009). The increase in latent heat transport is connected to the intense polar
 721 amplification of the EECO world. From the effected large scale circulation patterns, the Hadley
 722 cell changes both under the CO₂ and non-CO₂ constraints, with an asymmetric shift. More
 723 energy is being overturned in the northern cell, and less in the southern one, while the net
 724 poleward MOC heat transport stays close to the control simulation. Monsoon systems are also
 725 affected by both the non-CO₂ and CO₂ constraint, but in an opposite way. We found smaller
 726 monsoon areas in the Eocene, due to the different topography, while at a higher CO₂
 727 concentration, the warmer atmosphere's higher water holding capacity means that more heat is
 728 transported poleward by the monsoons. At the midlatitudes cyclones' heat transport increase
 729 mainly in the Northern Hemisphere, due to the before mentioned topography differences.

730 In summary we found that transport processes indicate a more intense hydrological cycle
 731 and also polar amplification in the warmer EECO compared to the present-day climate. We
 732 identified which processes are affected by the Eocene boundary conditions and which are

733 sensitive to the CO₂ increase. The different boundary conditions cause more southward transport
734 in the ocean which is only partially compensated by the atmosphere. The Eocene set up increases
735 the heat transport of northern midlatitude cyclones and decreases the global monsoon area. The
736 CO₂ increase, on the other hand causes an increase in poleward atmospheric heat transport.
737 Regarding the large scale circulation patterns the CO₂ increase results in an increase in the latent
738 heat transport of monsoon systems and more overturning heat in the Hadley circulation, although
739 the net poleward transport of the Hadley cells is stable. Nevertheless, there is an asymmetry
740 between the northern and southern Hadley cell, with larger changes in the northern cell. A more
741 detailed analysis of the largescale circulation patterns of the Eocene climate, in higher temporal
742 and spatial resolution model results, and their comparison to proxy data, is the focus of our
743 further research.

744

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746

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761

762 **Open Research**

763 The DeepMIP PI and Eocene simulations are available by following the instructions
764 at <https://www.deepmip.org/data-eocene/>; please see (Lunt et al., 2021). Hersbach, H. et al.,
765 (2019) was downloaded from the Copernicus Climate Change Service (C3S) Climate Data Store.

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