1 **Transition into a Hothouse World at the Permian-Triassic Boundary – A model**
2 **study**

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13 **Keywords:**

14 Permian–Triassic boundary

15 Mass extinction

16 Anoxia

17 Primary productivity

18 Clouds feedbacks

19 Global climate change

20 Climate modeling

21

Abstract. The Permian-Triassic Boundary (PTB, ~252.3 Ma) marks the largest mass extinction of the Phanerozoic, with a loss of more than 90% of marine organisms, and is characterized by lethally hot surface temperatures. The PTB global warming has been linked to greenhouse gas emissions from the Siberian Traps and associated coal-bed intrusions and likely led to severe environmental consequences, such as ocean acidification, a decline in the marine productivity, and extensive hypoxia. In order to understand these changes, feedbacks in the climate system have been explored with sensitivity climate simulations and compared to temperature proxies from the sedimentary record. The response of the PTB ocean circulation to an atmospheric perturbation of ~5,000 PgC, comparable to Earth's total fossil fuel inventory, leads to a global temperature increase by 3-4 °C and an increase in ocean stratification. The pole-to-equator temperature gradient decreases by 2 °C with an increase in CO₂-radiative forcing, predominately due to snow albedo feedbacks over Gondwana. The greenhouse-induced warming would have led to a weakening of the Hadley cell and an associated decrease in the trade winds and equatorial primary productivity. These climatic changes might have been amplified by cloud-feedback processes. A reduced concentration of cloud condensation nuclei due to a biologic decline of dimethylsulfide caused by the temperature stress or changes in airborne mineral particles could have reduced the cloud albedo, particularly in high latitudes. Results from a climate simulation with reduced cloud albedo suggest a polar warming of up to 7 °C and a reduction of the pole-to-equator temperature gradient by ~4 °C in addition to the reduction caused by the increase in greenhouse-induced radiative forcing. The scenario with reduced cloud albedo

44 further leads to an increase in ocean stratification and widespread low-oxygen
45 concentrations in the Panthalassa during the Early Triassic.

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1. Introduction

The climatic transition from the Late Permian to the Early Triassic is of great significance because it encompasses a transition into a hothouse world (Joachimski et al., 2012; Sun et al., 2012), widespread variation in anoxia (Cao et al., 2009; Hallam and Wignall, 1997; Isozaki, 1997; Knoll et al., 1996; Kump et al., 2005; Meyer et al., 2008; Shen et al., 2010; Ward et al., 2001; Wignall and Hallam, 1992, 1993; Wignall and Newton, 2003; Wignall and Twitchett, 2002), and ocean acidification (Black et al., 2014; Clarkson et al., 2015; Clapham and Payne, 2011; Fraiser and Bottjer, 2007a, 2007b; Heydari et al., 2008; Payne et al., 2010). The mass extinction at the Permian-Triassic Boundary (PTB) 252.3 Ma (Shen et al., 2011), affected over 90% of marine species (Alroy et al., 2008; Bambach et al., 2004; Clapham and Bottjer, 2007; Erwin, 1994; Knoll et al., 2007; Kozur, 1998; Raup, 1979; Sepkoski, 1995; Visscher et al., 2004) and was a major driver of macroevolutionary change in the early Mesozoic (Burgess et al., 2014). Common hypotheses for causes of the event include the massive volcanic eruption of the Siberian Traps (Bowring et al., 1998; Campbell et al., 1992; Ganino and Arndt, 2009; Kamo et al., 2003; Mundil et al., 2004; Reichow et al., 2009; Renne and Basu, 1991; Renne et al., 1995; Sobolev et al., 2011; Svensen et al., 2009; Yin et al., 1996; Wignall, 2007) and associated volatile releases (Black et al., 2012), a massive bolide impact (Basu et al., 2003; Becker et al., 2001), thermogenic methane emission from intrusions into the West Siberian Coal Basin (Payne and Kump, 2007; Retallack and Jahren, 2008; Svensen et al., 2009) as supported by coal fly ash deposits in lake deposits (Grasby et al., 2011) and marine organisms (Nestell et al., 2015), and extensive wild fires

(Hudspith et al., 2014). Whether there was gradual, one-step or multi-stepped carbon release to the atmosphere is still under debate (Payne and Clapham, 2012).

Here, we investigate how changes in the greenhouse gas concentration and cloud optical depth can alter the climate and thus the solubility and biological pump and lead to an associated decline in oxygenation of the ocean during the climatic transition from the end-Permian to the Early Triassic.

The environmental changes at the PTB likely affected cloud optical properties (Heavens et al., 2012). Cloud coverage may change in response to higher CO₂ via stratospheric circulation and water content (Kirk-Davidoff et al., 2002; Sloan et al., 1995) or changes in the aerosol forcing (Heavens et al., 2012), and this change may contribute to a substantially warmer high-latitude climate. Current climate models generally do not consider interactions between atmospheric chemistry and extreme environmental changes such as an extensive release of hydrogen sulfide (Lamarque et al., 2007; Kump et al., 2005) that could alter cloud properties, ultimately altering polar stratospheric clouds and enhancing the ozone hole.

Planetary albedo, the fraction of incoming shortwave radiation reflected back to space, is highly dependent on the concentration of cloud condensation nuclei (CCN), aerosol particles consisting typically of a mixture of soluble and insoluble compounds on which water vapor condenses. In areas of high concentrations of CCN, for example in polluted areas, the cloud optical thickness and thus the cloud albedo are high, and the amount of heavy precipitation decreases, whereas in regions with lower concentrations of CCN, like over remote oceans, cloud optical thickness and cloud albedo are low and heavy precipitation is more intense because of increased droplet size. Substantial

amounts of CCN originate from inorganic substances (such as sulphate, nitrate, ammonium, sea salt), organic compounds, black carbon from biomass burning, airborne mineral particles (mostly from arid regions) and primary biological aerosol particles, ranging from a few per cm^3 in polar regions to a few 100 per cm^3 in unpolluted subtropical areas (Hudson and Sun, 2002).

Biological production is a major factor in providing natural sources of CCN (Andreae, 2007) and thus can alter the climate substantially. In this paper we hypothesize, that a reduction of the CCN and thus cloud droplet concentration and increased droplet radius (Kiehl and Shields 2013; Kump and Pollard, 2008) either by reduced marine biological production, changes in sea salt concentration, and/or mineral dust composition (Retallack, 2011) would have contributed to the global warming at the PTB (Sun et al., 2012). For example, a transition into a hothouse due to an increase in the atmospheric CO_2 concentration would likely lead to reduced upwelling and stratification at high latitudes (Kiehl and Shields, 2005) and to reduced productivity (Winguth and Winguth, 2012). Decrease in marine productivity in turn would potentially lead to a decrease in the emission of dimethyl sulfide (DMS) or dimethylsulfoniopropionate (DMSP) into the atmosphere, and a reduced DMS concentration in the atmosphere (Charlson et al., 1987; Kump and Lovelock, 1995; Lovelock and Kump, 1994). DMS eventually oxidizes to SO_2 and further to SO_4^{2-} . Thus reduced DMS concentrations will ultimately lead to reduced concentrations of sulfate CCN and a reduction in cloud optical depth, and hence exacerbate global warming.

2. Climate model and experiments

2.1 The Community Climate System Model (CCSM3)

The climate simulations for the Permian-Triassic transition have been carried out with the NCAR CCSM3, a global coupled climate model including the ocean, atmosphere, sea ice, and land surface (Collins et al., 2006a; Yeager et al., 2006; Kiehl and Shields, 2005; and references therein). A spectral horizontal resolution of T31, corresponding to an equivalent grid spacing of approximately $3.75^\circ \times 3.75^\circ$ in latitude and longitude, and vertically 26 unevenly spaced terrain-following levels (Collins et al., 2006b), has been chosen for the atmosphere to allow long-term integrations. Correspondingly, the ocean is resolved by a nominal 3° horizontal grid and by 25 layers in the vertical direction (Gent et al., 2006).

The marine carbon component of the coupled climate-carbon cycle model is based upon the Ocean Carbon Model Intercomparison Project (OCMIP) (Doney et al., 2006; Najjar and Orr, 1999; <http://www.ipsl.jussieu.fr/OCMIP/>) and is described in Winguth and Winguth (2013) for the Permian application. The model estimates air-sea fluxes of CO_2 with the wind-dependent gas exchange coefficient (Wanninkhof, 1992) across the air-sea interface, the temperature-dependent solubility of CO_2 , and the difference between the prescribed pCO_2 of the atmosphere and the pCO_2 of the uppermost layer of the ocean (Doney et al., 2006). The ocean carbon cycle model includes seven prognostic variables, i.e. phosphate (PO_4), total dissolved inorganic iron, dissolved organic phosphorus and iron, dissolved inorganic carbon, total alkalinity, and dissolved oxygen, that are transported by the ocean model. The parameterization of biological uptake of nutrients is similar to that used in the Hamburg Model of the Ocean Carbon Cycle (HAMOCC; Maier-Reimer, 1993) and assumes a constant Redfield ratio for particulate organic

matter. The uptake of PO_4 is given by the turnover of biomass, modulated by surface solar irradiance, temperature, and macro- and micronutrients. The carbon cycle model uses a single Martin power-law curve to describe the vertical particulate organic phosphorus flux profile over the full water column.

2.2 Permian boundary conditions and experiments

The Permian paleogeography, orbital settings, and solar constant (1338 W m^{-2} , $\sim 2.1\%$ lower than the modern value) applied for the Permian, are the same as used in Kiehl and Shields (2005). The effect of eccentricity-modulated changes of the precession on the strength of Pangean megamonsoons (Kutzbach and Gallimore, 1989) likely impacted export production and oxygen distribution (Winguth and Winguth, 2013). The prescribed land vegetation types are taken from Kiehl and Shields (2005) and based on surface types from Rees et al. (1999). Present-day conditions are applied for all other boundary conditions and parameterizations.

Two greenhouse gas levels were considered in this study, a $4\times\text{CO}_2$ simulation, four times higher than the preindustrial level (PAL) of 280 ppmv, and the $12.7\times\text{CO}_2$ (or $10\times\text{CO}_2$ relative to a reference value of 355 ppmv) simulation using the same atmospheric CO_2 concentration as Kiehl and Shields (2005). The first experiment follows the atmospheric pCO_2 estimates for the time prior to the onset of the Siberian Trap volcanism, whereas the 3550 ppmv for atmospheric pCO_2 (corresponding to a radiative forcing of 16 W m^{-2}), is assumed for an experiment simulating end-Permian conditions (based on Kidder and Worsley, 2004), as inferred from soil carbonate CO_2 paleobarometers (Cerling, 1991). Other greenhouse gas concentrations were set to 0.7 ppmv for pCH_4 and 0.275 for pN_2O .

The reconstructions of greenhouse gas concentrations at the end-Permian remain tentative because of the uncertain amount of carbon injection due to Siberian Trap volcanism-induced metamorphism of organic matter and petroleum (Svensen et al., 2009). The maximum end-Permian greenhouse gas concentration of ~12-13x PAL, as inferred from soil carbonate CO₂paleobarometers (Cerling, 1991), and consistent with estimates by Kidder and Worsley (2004), has recently been challenged by Breecker et al. (2010) who propose a lower end-Permian value of ~4–5x PAL, in agreement with the GEOCARBSULF model (Berner, 2006) and Montanez et al. (2007), inferred from Middle Permian goethites from the eastern shelf of the Midland Basin. Such low estimates, however, are inconsistent with emission estimates from the Siberian Traps and successive Late Permian and Early Triassic spikes of unusually high atmospheric CO₂ (Payne and Kump, 2007; Payne et al., 2004) inferred from stomatal index from fossil *Lepidopteris* leaves (Retallack et al., 2009, 2011), and are also in disagreement with the frost-free polar climates inferred from sensitive sedimentary deposits (Osen, 2014; Ziegler et al., 1998).

A third hothouse climate sensitivity experiment with reduced cloud droplet concentration at 12.7xCO₂PAL, named hereafter 12xCO₂ Warm Climate (WC), addresses the role of changes in cloud albedo in response to changes in the cloud droplet concentration, as discussed in section 1. Here we follow the approach of Pollard and Kump (2008; see also Kiehl and Shields, 2013) that a reduction in past CCN concentrations has a profound effect on both the cloud droplet concentration and the cloud droplet radii. As a guidance we select present-day anthropogenically undisturbed regions with cloud droplet concentrations of 50 cm⁻³ and cloud radii (r_{liq}) of 17 μm (Table 1).

Such a change would result in reduction of cold cloud droplet concentration over the ocean by 67%, a decrease of warm continental cloud droplet concentration by 88%, as well as increases in marine and continental liquid droplet radii by ~21% and 113%, respectively, leading to a decrease of the pole-to-equator temperature gradient and global warming. These possible changes reflect the maximum likely effect of extremely reduced CCN production rates on global warmth, which could have contributed to the hothouse conditions in the Early Triassic.

All simulations have been integrated for ~3,000 years into quasi-steady state from the ~5,000-year integration of the 12.7xCO₂ simulation and the results are discussed in the following sections.

3. Results

3.1 Cloud cover and precipitation

In this section, vertical changes in cloud fraction and total cloud coverage are examined. The present-day CCSM3 simulation (Table 2) generally agrees with the observations except for tropical cloud fraction at 100 mb in the atmospheric component CAM3 (Collins et al., 2006b), but potentially underestimates subvisible clouds (Boville et al., 2006).

For PTB boundary conditions, two experiments are considered, one with reduced cloud droplet concentration and increased droplet size, the 12.7x CO₂ WC experiment, and the PTB 12.7x CO₂ reference scenario of Kiehl and Shields (2005). In general, a reduction in low cloud coverage leads to warming because of the lower albedo in cloud-free areas. On the other hand, an increase in high cloud concentration, particularly

in thin cirrus clouds, contributes to warming, because these clouds are transparent to incoming short-wave radiation and absorb outgoing long-wave radiation. The meridional vertical section of cloud fraction (Fig.1) displays a lower cloud fraction in the lower troposphere in the 12.7x CO₂ WC relative to the 12.7x CO₂ simulation in the westerlies zone (around 60°) between 850 mb and 700 mb, particularly in the northern hemisphere, where cloud fraction is lowered by 2-7% due to the reduction of cold CCN over the ocean (Table 2). On the other hand, cloud fraction increases at these latitudes by a similar magnitude above 200 mb. Tropical cloud fraction in the 12.7x CO₂ WC experiment exceeds the one in the PTB reference scenario by 4% between 850 mb and 300 mb, whereas cloud fraction decreases just above this height, related to a decrease in the ice cloud particle density relative to the reference 12.7x CO₂ simulation.

Overall, total cloud cover (Fig. 2a and b) decreases over land, particularly at 30° and 60°, whereas over the central equatorial Panthalassa there is an increase of convective activity and cloud coverage by 12%. This increase in convective activity is also supported by enhanced precipitation of >5 mm day⁻¹ in the 12.7x CO₂ WC experiment compared to the 12.7x CO₂ scenario. Global rainfall in this scenario is 27% higher (Table 2) than for the corresponding reference simulation because of enhanced rainfall over land due to reduced cloud droplet concentration and enhanced cloud droplet radii and over the tropical ocean related to the enhanced convective activity as discussed above. Interestingly, these differences in rainfall over land and ocean (Fig. 3 a and b) linked to changes in the cloud droplet parameters and the associated cloud albedo are higher compared to rainfall increase due to atmospheric greenhouse gas concentration changes from 4x CO₂ PAL to 12.7x CO₂ PAL (equivalent to a ~5000 Pg C change or ~1/6 of the total release of 30,000

Pg C estimated by Svenson et al., 2009; Table 2). Global precipitation increases by 11% in to response to this carbon pulse, partly due to enhanced tropical convection over the warm pool of the eastern equatorial Panthalassa and partly linked to an increase in precipitation at high latitudes which is consistent to future climate predictions with CCSM3 (Meehl et al., 2006).

3.2 Climate sensitivity to changes in greenhouse gas concentration and cloud cover

In this section, the climate sensitivity to greenhouse gas forcing across the PTB is discussed for the two different greenhouse gas scenarios, the 4x CO₂ and 12.7x CO₂ PAL experiments, and for the 12.7x CO₂ WC simulation with altered cloud droplet parameters (Table 1). Figure 4 displays computed temperature changes (ΔT) as a function of various climate sensitivities for a doubling of the atmospheric CO₂ ($\lambda = 1.5, 3.0, 4.5, 6.0$) and the radiative forcing (RF_{CO_2}):

$$\Delta T = \lambda RF_{CO_2}. \quad (1)$$

For a given atmospheric CO₂ concentration, radiative forcing (RF_{CO_2}) in eq. (1) is computed after Hansen et al. (1988)

$$RF_{CO_2} = 6.3 \ln \left(\frac{pCO_2}{pCO_{2(ref)}} \right) \quad (2)$$

with a reference value for pCO_{2(ref)} of 280 ppmv.

Also shown is the difference of simulated temperature between the sensitivity experiments (T_{PTB}) and the preindustrial simulation T_{PI} (Otto-Bliesner et al., 1996) with a global mean surface temperature of 12.8 °C

$$\Delta T_{model} = T_{PTB} - T_{PI} + \Delta T_S \quad (3)$$

where a temperature correction $\Delta T_S = \sqrt[4]{\frac{S_{PI}(1-\alpha)}{4\epsilon\sigma}} - \sqrt[4]{\frac{S_{PTB}(1-\alpha)}{4\epsilon\sigma}} = 1.4^\circ\text{C}$ has been applied based on an energy balance model (North et al., 1983) to consider a 2.1% reduction of the solar constant (S) relative to the preindustrial (PI) conditions of 1365 W m^{-2} . Here, global albedo is approximated to be 30%, a global emissivity of $\epsilon = 0.61$ is assumed, and the Stefan-Boltzmann constant is set to $\sigma = 5.67 \cdot 10^{-8}\text{ W m}^{-2}\text{ K}^{-4}$.

The climate sensitivity of CCSM3 for the PTB with this temperature correction is $\sim 3.2^\circ\text{C}$ in response to a doubling of the atmospheric CO_2 concentration (as inferred from eq. 1 and Fig. 4) for the scenario with standard cloud parameterization in agreement with the IPCC 2013 report, whereas a higher climate sensitivity of $\sim 4.6^\circ\text{C}$ is achieved by the $12.7\times\text{CO}_2$ WC simulation with a reduced cloud albedo.

3.3 Sensitivity of the equator-to-pole gradient and temperature pattern to changes in greenhouse gas forcing and cloud parameterization

Polar amplification and associated reductions in latitudinal temperature gradients have been an omnipresent feature of climate model simulations with increased CO_2 (Manabe et al., 1991), but the mechanisms that drove the polar warmth at the PTB are controversial. Hothouse worlds, like the Early Triassic are generally characterized by lower-than-present pole-to-equator gradients, however, past climate model simulations so far have generally underestimated polar temperatures in hothouse climates (e.g. Barron and Washington, 1982; Gibbs et al., 2002; Hotinski et al., 2001; Winguth et al., 2002; Lunt et al., 2012; Zhang et al., 2001), as e.g. demonstrated by discrepancies between climate simulation results and $\delta^{18}\text{O}$ analyses or reconstructed biomes (Osen, 2014; Rees et al., 1999). Clouds have a substantial influence on polar energy budgets of past climates

(e.g., Kiehl and Shield, 2013; Sloan and Pollard, 1998; Upchurch et al., 2015) and temperature amplification (Winton, 2006; Kay et al. 2012).

Our pole-to-equator temperature distributions(Fig.5) show the transition from a 4xCO₂ climate (during the middle Permian) to a 12.7x CO₂ climate (at the PTB). The simulated tropical sea surface temperature for pre-PTB conditions of about ~29.3 °C in the equatorial Tethys Ocean (Fig. 6a) is in agreement with temperature reconstructions from oxygen isotope measurements in conodonts and brachiopods for the South China Shangsi Section (20 °N) (Joachimski et al., 2012; Sun et al., 2012) and the two sections in NW Iran (Kuh-e-Ali Bashi and Zal) near the paleo-equator (Schobben et al., 2014). This value is about 4 °C higher than the zonal average in Figure 5, predominately related to the cold tongue of the eastern equatorial upwelling of colder subsurface water masses in the Panthalassa. Northern high-latitude temperatures are above freezing with ~4 °C mostly due to heat flux from the ocean whereas in the southern hemisphere, the landmasses of Gondwana lead to a mean temperature below freezing between 68°S and 85°S (Fig. 5). The difference in radiative forcing between the 12.7x CO₂ simulation and the 4x CO₂ simulation results in a global mean difference of 4.2°C (Table 2) and a zonal mean temperature difference of 3 °C in the tropics, whereas polar temperatures are amplified by 4°C in the northern hemisphere and 4.5°C in the southern hemisphere.

The corresponding variations in temperature patterns(Fig.7) follow two main distinct characteristics of future temperature predictions(IPCC,2013): firstly, the temperature over land increases more rapidly than over the ocean due to different local feedbacks, differences in latent heat flux, differences in heat capacity between the land and ocean, and thermal inertia (IPCC, 2013; Manabe et al., 1991; Sutton et al., 2007), and

secondly, polar temperature changes are amplified related to snow and sea ice albedo feedbacks. Note that the global coverage of sea ice (Table 2) in the 4x CO₂ experiment is <1%, so that changes due to the sea ice albedo feedback are relatively small. Maximal temperature differences of up to 9 °C are simulated over the polar region of Gondwana due to snow albedo feedbacks (note that maximal water equivalent snowdepth is <1 cm for the annual average in all simulations).

Our simulation with reduced cloud albedo (12.7x CO₂ WC) predicts a remarkable 4.5 °C higher global average surface temperature than the 12.7x CO₂ simulation of Kiehl of Shields (2005). Annual average temperatures exceeding 35°C in the equatorial Tethys are in the range of uncertainty of the oxygen isotope-based temperature reconstructions from conodonts for the Early Triassic from the Iranian and the South China sections (Joachimski et al., 2012; Schoppen et al., 2014; Sun et al., 2012; Figs. 6 and 7). Relative to the 12.7xCO₂ experiment, tropical temperature increases by 4 °C, with the highest temperature of 38.2°C in January near South China (Fig. 7c).

Reduced tropospheric cloud cover in mid-to-high latitudes (>40°; Fig. 2) is predicted by the climate scenario with reduced cloud albedo. Such a reduction in cloud optical depth leads to a remarkable polar surface temperature amplification (>5 °C). Lower cloud cover over the land (Fig. 2b) in the 12.7x CO₂ WC causes a substantial increase in temperature, particularly during southern hemisphere winter, where reduced snow cover leads to a positive feedback. Extreme environmental conditions are simulated for Gondwana near 30°S, with annual average temperatures exceeding 50 °C (or for comparison about 25 °C higher than Death Valley, California, U.S.A.). Such extreme temperature conditions affect the stability of protein structures (Somero, 1995) and thus

survival of higher organisms.

The vertical temperature structure (Fig. 6a) along the equator from 5 °S to 5 °N in the 4xCO₂ experiment shows the east-to-west gradient of the thermocline with shallower depth in the eastern equatorial Panthalassa associated with upwelling of subsurface water masses. Note that the temperature perturbation near 90 °E is induced by topographic features near South China. Differences between the 12.7x CO₂ scenario and the 4xCO₂ scenario are characterized by a positive vertical density gradient anomaly for surface-to-intermediate water masses. This shift is linked to an increase in thermal stratification by ~2°C (Fig. 6b) in response to an increase in CO₂-radiative forcing and amplified by an increase in the vertical halocline in response to surface freshening, in particular near the surface in the western Panthalassa and Tethys Ocean. A positive shift of 4°C is simulated for the 12.7xCO₂ WC scenario (Fig. 6c) relative to the baseline 12.7xCO₂ experiment with a comparable higher vertical thermal gradient and thus a positive vertical halocline anomaly, leading to reduced vertical mixing. Enhanced stratification and ocean warming in the 12.7xCO₂ WC scenario has implications for the solubility and biological pump. An increase in temperature reduces the saturation level of dissolved gases such as O₂ and CO₂. The latter leads to a reduction in the uptake of dissolved CO₂ in seawater and thus a positive climate feedback. Increased stratification also leads to an enhanced biological pump, and thus stronger vertical gradients in dissolved oxygen and total dissolved inorganic carbon.

3.4 Export production

Environmental changes due to changes in radiative forcing and cloud albedo lead to

variations in the export production (Fig. 8) which depends on macro- and micronutrient supply, temperature, available light and mixing. For example, according to the temperature parameterization for export production used in this study (Maier-Reimer, 1993), a 5 °C increase of the temperature to 30 °C enhances the export production by 3%.

In the equatorial Panthalassa, the weakening of trade winds with an increase in CO₂ radiative forcing induces a decline of Ekman-induced upwelling, thus reducing the supply of nutrient-rich subsurface water masses. A reduction of upwelling leads to surface warming (Figs. 6 and 7) and temperature-induced enhanced productivity, but the lack of nutrients due to reduced upwelling appears to be the dominant factor in this region, reducing the maximum equatorial export production by 47% from 15.6 mol m⁻² yr⁻¹ in the 4x CO₂ scenario to 8.3 mol m⁻² yr⁻¹ in the 12.7x CO₂ scenario (Fig. 8b). The simulated differences between the 4x CO₂ and 12.7x CO₂ scenario at high latitudes are of more complex nature and linked to changes in the Fe-cycle and temperature. Simulated export production at 60 °N is affected by an increase in the Fe concentration with higher CO₂-radiative forcing that is linked to a shift in the barotropic stream function. Note that although PO₄ concentration is reduced in the 12.7x CO₂ scenario at this latitude, the absolute PO₄ concentration ranging from 0.8 μmol L⁻¹ to 1.8 μmol L⁻¹ is not limiting the export production. Higher temperature in the 12.7x CO₂ simulation (Fig. 6b) further stimulates the export production. Similarly, an increase in high-latitude export production in the southern hemisphere is also linked to an increase in the surface Fe concentration and temperature.

The changes in cloud properties further alter the pattern of export production. Figures 8a and 8c show the pattern of export production for the 12.7x CO₂ WC scenario.

The general feature is a weak export production with maximum values of $\sim 12 \text{ mol C m}^{-2} \text{ yr}^{-1}$. In the eastern equatorial Panthalassa, there is a $\sim 3 \text{ mol C m}^{-2} \text{ yr}^{-1}$ increase in the export production (Fig 8c) related to a $\sim 4^\circ \text{C}$ increase of temperature compared to the reference 12.7x CO_2 experiment (Winguth and Winguth, 2013) albeit a slightly reduced Ekman-induced upwelling. In contrast, export production in the western part of the equatorial Panthalassa is reduced by $\sim 1 \text{ mol C m}^{-2} \text{ yr}^{-1}$ associated with the strengthening of the trade winds. Interestingly, export production in subpolar regions near 60° decreases because of less available Fe in the region. The lower Fe concentration is associated with increased stratification and the associated enhanced nutrient gradient.

3.5 Dissolved Oxygen

The distribution of dissolved oxygen in the ocean depends on solubility, the biological pump and ocean circulation. For example, solubility of dissolved oxygen decreases by 10% from 4.4 ml L^{-1} at 30°C , $S=35$ and $P=1013 \text{ hPa}$ to $\sim 4.0 \text{ ml L}^{-1}$ at 35°C (Weiss, 1970). The simulated distribution of dissolved oxygen along an east-west equatorial section of the Panthalassa and Tethys Ocean in the 4x CO_2 scenario (Fig. 9a), as an analog for pre-PTB conditions, is generally comparable with observed present-day conditions in the Pacific Ocean with a core of the oxygen minimum zone (OMZ) at $\sim 400\text{--}500 \text{ m}$ as inferred from the World Ocean Circulation Experiment (Wijffels et al., 1996). Here, we define the OMZ as the region where the dissolved oxygen concentration $[\text{O}_2]$ is hypoxic ($0.0 \text{ ml L}^{-1} < [\text{O}_2] \leq 1.4 \text{ ml L}^{-1}$). The effective biological pump is similar to its present strength; equatorial upwelling and export production result in an effective export of POC into the twilight zone, where dissolved oxygen is depleted by remineralization

of POC.

Interestingly, the carbon pulse leading to an atmospheric CO₂ concentration of 12.7x CO₂ corresponding to the mid- to end-Permian transition results in the decline of the areal extent of the OMZ by ~25%, because of the reduced export production (Fig 8b). Further global warming comparable to the transition into the Early Triassic as simulated by reduced cloud optical depth in the 12.7xCO₂ WC scenario leads to an expansion of the OMZ by ~40% compared to the 12.7xCO₂ experiment, linked to enhanced stratification (Fig. 7c), an increase in export production in the eastern equatorial region, and an increase in temperature of the water masses. Note that in this scenario, 47% of the equatorial Panthalassa in 500 m depth is hypoxic and would favor the formation of black shale sediments at that depth, in particular in the eastern part where anoxic conditions ([O₂] ≤ 0.0 ml L⁻¹) are simulated.

4. Discussion

Climate simulations presented in this paper demonstrate potential feedbacks in the climate system (Fig. 10) during the transition into a hothouse climate at the PTB. Siberian Traps volcanism (Campbell et al., 1992; Bowring et al., 1998; Ganino and Arndt, 2009; Kamo et al., 2003; Reichow et al., 2009; Renne and Basu, 1991; Renne et al., 1995; Svensen et al., 2009; Wignall, 2007; Yin et al., 1996), coal bed intrusions (Retallack and Jahren, 2008), and forest fires (Hudspeth et al., 2006; Uhl et al., 2007; Wang and Chen, 2001; Wang et al., 1996) supported by coal fly-ash deposits (Nestell et al., 2015; Grasby et al., 2011) would have led to substantial carbon emissions (Brand et al. 2012; Retallack et al., 2011; Svensen 2009), and resulted in climatic warming (Hallam and Wignall, 1997;

Joachimski et al., 2012; Kidder and Worsley, 2010; Sun et al., 2012; Schoppen et al., 2014). The resulting extensive loss of forests and marine organisms would have reduced carbon burial on land and at the sea floor, thus potentially leading to a small amplification of greenhouse gases in the early Triassic (Berner, 1991; Broecker and Peacock, 1999). If we assume that the carbon inventory of land biota was ~2000 PgC (comparable to the present) and that this amount would have been emitted into the atmosphere, the long-term increase in atmospheric $p\text{CO}_2$ would have amounted to ~70 ppmv when taking into account carbon uptake by the ocean (Berner, 2002). Mass mortality in the ocean (and hence cessation of the biological pump) may result in an increase of the atmospheric $p\text{CO}_2$ of ~200-300 ppmv (Maier-Reimer et al., 1996; Berner, 2002). On longer time scales, silicate weathering can potentially compensate for the combined effects of these atmosphere-ocean carbon releases (Archer et al., 1999).

There is a substantial difference between the temperature increase of 3-4 °C in the shallow tropics from the 4x CO_2 to the 12.7x CO_2 simulation (equivalent to an atmospheric increase of 5000 Pg C) and the ~8 °C warming in the Tethys Ocean inferred by Joachimski et al. (2012). This discrepancy can be attributed to uncertainties in both modeling and/or data. Higher values than the assumed 5000 Pg C increase in the atmosphere may have been possible for the Early Triassic (e.g. Svenson et al., 2009). For example, an increase in atmospheric $p\text{CO}_2$ equivalent to 20x the preindustrial value would, according to eq (1), result in a global temperature change of 10.8 °C relative to preindustrial conditions or 1.6 °C relative to the 12.7x CO_2 PTB simulation. In addition, changes in cloud aerosol forcing (Heaven et al., 2013) and improvements of model physics (Gent et al., 2014) could potentially raise tropical temperature by 4-5 °C relative

to the 12.7xCO₂ PTB simulation with CCSM3. Model resolution is also of significance; a higher model resolution would improve the representation of the ITCZ and of small-scale features (e.g. Small et al., 2014; Yeager et al., 2006). Uncertainties in the temperature reconstructions from oxygen isotopes include the assumption of the Vienna Standard Mean Ocean Water value of -1‰ for Late Permian seawater, no secular variations in seawater salinity, and variations in isotope fractionation among different taxa (Joachimski et al., 2012). Data sampling may also be biased towards certain seasons and certain times during longer-term climate variations such as orbital cycles (Winguth and Winguth (2013).

We focus here on a positive short-term climate feedback that is associated with the increase in greenhouse gas concentrations and reduced marine productivity, and possibly amplified Early Triassic warming by a decline in the concentration of cloud droplets and increased cloud droplet radii. Diminished productivity by global warming on the short geologic time scale (<5,000 yrs) reduces productivity, thus lowering DMS and DMSP emissions from e.g. green algae (Van Alsyne et al., 2003) and *Phaeocystis* species (Falkowski and Knoll, 2007; Schoemann et al., 2004) or other micro-organisms. DMS emissions may be also enhanced by ocean acidification (Six et al., 2013). The resulting reduction in sulfur aerosols and CCN concentrations can lead to enhanced cloud droplet size and reduced cloud albedo and cloud optical depth. In addition to these changes, other variations in CCN need to be considered to account for the shift in cloud optical properties as simulated in this study. For example, the composition of airborne mineral particles in air masses across the PTB (Heavens et al., 2012; Retallack et al., 2011), shifts in marine ecosystems (Broecker and Peacock, 1999; Knoll et al. 1996; Payne and

Chapman, 2012) or enhanced H₂S emissions (Meyer et al., 2008) could have altered the cloud properties and therefore produced global-scale changes in atmospheric circulation (Federov et al., 2013; Upchurch et al., 2015). The lower cloud optical depth as simulated in this study leads to a global average temperature increase of 4.5 °C, with extreme regional temperatures of >38 °C in the Tethys Ocean (Joachimski et al., 2012; 2014) and >50 °C on land and an increase in vertical stratification in the ocean, resulting in an expansion of the oxygen minimum zone and low oxygen environments (Hallam and Wignall, 1997; Isosaki et al., 1997; Knoll et al., 1996; Ward et al., 2001; Wignall et al., 1996; Wignall and Hallam, 1992, 1993; Wignall and Newton, 2003; Wignall and Twitchett, 2002). The processes described above act on relatively short timescales, with low productivity during the early Triassic (in the aftermath of the PTB extinction) resulting in reduced DMS emissions, thus favoring lower CCN concentrations and hothouse conditions during that time. On longer timescales, however, an increase in the marine nutrient inventory due to weathering intensification by global warming, increased rain fall, and acid rain (Black et al., 2014), followed by soil erosion (Sephton et al., 2005) that is also enhanced by forest fires (Hudspeth et al., 2014), would ultimately lead to enhanced marine productivity (Algeo et al., 2011) and anoxia (Meyer et al., 2008; Winguth and Winguth, 2012; Grasby et al., 2013). An associated rise in DMS emissions from the ocean and the resulting increase in cloud albedo by higher CCN concentrations would counteract the warming and weathering (Fig. 10).

The simulated tropical oceanic conditions in the Early Triassic are favorable to the formation of intense tropical cyclones (Emanuel et al., 2008; Kidder and Worsley, 2010) under the condition that the vertical shear of horizontal winds is low

(Knutson et al., 2010). These tropical cyclones would ultimately enhance vertical mixing of surface water masses, thus contributing to surface cooling and stronger poleward heat transport (Korty et al., 2008).

Regional temperature distributions are further affected by orbital pacing inferred from cyclical lacustrine rocks (Olsen and Kent, 1996) that could intensify the monsoon circulation in the Tethys Ocean (Crowley et al., 1992; Winguth and Winguth, 2013).

5. Conclusions

End-Permian greenhouse gas release from Siberian Trap volcanism and associated thermogenic methane release due to coal intrusions likely triggered climatic changes that were enhanced by positive feedbacks, ultimately leading to hothouse conditions and a substantial expansion of the oxygen minimum zone in the Early Triassic. Enhancement of the Early Triassic warming potentially occurred as a result of the decline in marine productivity in the aftermath of the end-Permian extinction, and thus reduced DMS and DMSP emission into the atmosphere. The decrease in sulfate aerosol particle concentrations that can act as cloud condensation nuclei progressively lowered the cloud albedo and enhanced the surface warming, particularly in polar regions. Simulated extreme Early Triassic hothouse conditions produced by such a mechanism are corroborated by the $\delta^{18}\text{O}$ analysis of conodonts and brachiopods.

Even an atmospheric carbon inventory change of ~5,000 Pg C (which is by far lower than estimates by Svenson et al., 2009) could lead to unsustainable conditions for life in large land areas due to the expansion of desert environments, and in the ocean due to ocean acidification (Clapham and Payne, 2011; Fraiser and Bottjer, 2007; Heydari et al.,

2008; Payne and Kump, 2007; Payne et al., 2010) and the associated decline of reefs (Kiessling and Simpson, 2011), among other inferred environmental changes (see e.g. review by Wignall, 2007).

There is a need to consider all potential feedback mechanisms between marine productivity, cloud optics, and the climate (e.g. Cameron-Smith et al., 2011; Gunson et al., 2006; Kloster et al., 2007; Six et al., 2013) in earth system modeling studies to improve the hindcast of past climates as well as future predictions. This study provides new insights into the role of clouds for a transition into a hothouse world. The paleorecord needs to be extended to identify potential sources and sinks of aerosols and to advance the interpretation of paleoclimatic changes during the Permian-Triassic transition.

Acknowledgements

We thank Jeff Kiehl at NCAR for discussing the boundary conditions for these simulations. We are also grateful to the two anonymous reviewers for their comments, and to Angela Osen for assistance in graphical programs. All model simulations were done on NCAR and NWSC computers, supported by NSF. The work is supported by NSF grant EAR-0745817.

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909

Figure 1

Figure 2

Figure 3

Precipitation [mm/day]

a)

$12.7 \times \text{CO}_2 \text{ WC}$

b)

$12.7 \times \text{CO}_2 \text{ WC minus } 12.7 \times \text{CO}_2$

Figure 4

Figure 5

Figure 6

Figure 7

Surface Air Temperature [°C]

a) $12.7 \times \text{CO}_2$ WC

b) $12.7 \times \text{CO}_2$ minus $4 \times \text{CO}_2$

c) $12.7 \times \text{CO}_2$ WC minus $12.7 \times \text{CO}_2$

Figure 8

Export Production [$\text{mol m}^{-2} \text{yr}^{-1}$]
and Vertical Velocity [10^{-6} m s^{-1}]

a) $12.7 \times \text{CO}_2 \text{ WC}$

b) $12.7 \times \text{CO}_2$ minus $4 \times \text{CO}_2$

c) $12.7 \times \text{CO}_2 \text{ WC}$ minus $12.7 \times \text{CO}_2$

Figure 9

Figure 10

Table 1

Experiments	Atm. CO ₂ Conc. (ppmv)	Direct CO ₂ Rad. Forcing (W m ⁻²)	CDNC* Warm Cloud (cm ⁻³)	CDNC Cold Cloud (cm ⁻³)	CDNC Ice Cloud (cm ⁻³)	r _{liq} Ocean (μm)	r _{liq} Land (μm)	r _{liq} Ice (μm)
Present Day (PD)	355	1.5	400	150	40	14	8	14
PTB 4xCO ₂	1120	8.7	400	150	40	14	8	14
PTB 12.7xCO ₂	3556	16.0	400	150	40	14	8	14
PTB12.7xCO ₂ WC	3556	16.0	50	50	50	17	17	17

* Cloud Droplet Number Concentration

Table 2

Experiments	SAT	SAT	SST	Precip.	Precip.	Precip.	Low	A _{ice} (%)
	Global	Land	(°C)	Global	Land	Ocean	Cloud	
	(°C)	(°C)		(mm/day)	(mm/day)	(mm/day)	Cover	
							(%)	
Present Day (PD)	14.1	17.8	19.4	2.7	2.0	3.2	43.3	26.6
PTB 4xCO ₂	18.2	15.1	19.3	2.8	1.8	3.2	39.4	0.8
PTB 12.7xCO ₂	22.0	20.0	22.7	3.0	2.0	3.5	39.1	0.0
PTB12.7xCO ₂ WC	26.5	26.2	26.5	3.8	2.4	3.7	38.3	0.0

Table Captions

Table 1. List of experiments and important CCSM3 parameters altered in this study.

Table 2. Global averages from the CCSM3 simulations.

Figure Captions

Fig. 1. Zonal averaged cloud fraction (%) for the PTB 12.7 x CO₂ WC simulation with CCSM3 (a) with reduced cloud droplet concentration and increased droplet radii (Table 1) and difference between this simulation and the reference simulation 12.7 x CO₂ of Kiehl and Shield (2005).

Fig. 2. Total cloud cover (%) for for the PTB 12.7 x CO₂ WC simulation with CCSM3 (a) with reduced cloud droplet concentration and increased droplet radii and difference between this simulation and the reference simulation 12.7 x CO₂ of Kiehl and Shield (2005).

Fig. 3. Total precipitation (mm day⁻¹) for for the PTB 12.7 x CO₂ WC simulation with CCSM3 (a) with reduced cloud droplet concentration and increase droplet radii and difference between this simulation and the reference simulation 12.7 x CO₂ of Kiehl and Shield (2005).

Fig. 4. Temperature change as a function of various climate sensitivities ($\lambda=1.5$, blue, 3.0, black, 4.5, red, 6.0, purple) to a doubling of CO₂ radiative forcing based on Hansen et al., 1988 (eq. 1-2) and temperature difference between the PTB and preindustrial simulations of CCSM3 as a function of the atmospheric CO₂ concentration relative to pre-industrial levels (eq. 3). Solid stars denotes PTB radiative forcing and unfilled star consider a temperature correction $\Delta T_S = 1.4$ °C due to the 2.1% reduction of the solar constant for the PTB relative to present day using an energy balance model whereas non-filled stars display the temperature change without the temperature correction. Climate simulations without present-day cloud optical depth (blue) for

and with reduced cloud droplet concentration and increase droplet radii (12.7 CO₂ WC; red) relative to the preindustrial simulation (Otto-Bliesner et al., 2006).

Fig. 5. Zonal average surface air temperature for CCSM3 4 x CO₂, 12.7 x CO₂, and 12.7 CO₂ WC climate simulations. For comparison temperature reconstruction derived from oxygen isotope measurements of conodonts and brachiopods samples for the mid latitudes (Mei and Henderson; 2001; green blue stars); for South China at ~20° N (Joachimski et al., 2012) for the Late Permian (Clarkina sp. and Hindeodus sp., rectangles white) and early Triassic (Hindeodus sp., blue rectangles); and for NW Iran near the equator (Schoppen et al., 2014) for the Late Permian (Clarkina sp., white circles, and brachiopods, black circles) and Early Triassic (Hindeodus sp., and indet., grey diamonds, and Hindeodus sp., white diamonds).

Fig. 6. Vertical equatorial section of ocean temperature of CCSM3 for the PTB for the (a) 4 x CO₂ experiment, (b) 12.7 x CO₂, and (c) 12.7 x CO₂ WC experiment. Contour interval is 1 °C. Note that the left ocean basin is the Tethys Ocean and right ocean basin denotes the Panthalassa. Temperature between 30 °C and 35 °C are light gray shaded and >35 °C are dark gray shaded.

Fig. 7. Surface air temperature at the PTB for (a) the 12.7 x CO₂ WC simulation with CCSM3, (b) differences between the 12.7 x CO₂ and 4 x CO₂ experiment, and (c) differences between the 12.7 x CO₂ and 4 x CO₂ experiment. Contour interval is 5 °C.

Fig. 8. Export production and vertical upwelling at the base of the euphotic zone at the PTB for (a) the 12.7 x CO₂ WC simulation with CCSM3, (b) differences between the 12.7 x CO₂ and 4 x

CO₂ experiment, and (c) differences between the 12.7 x CO₂ and 4 x CO₂ experiment. Contour interval is 1 mol m⁻² yr⁻¹ for export production and 0.5 10⁻⁶ m s⁻¹ for vertical velocity. Dashed lines denote upwelling.

Fig. 9. Vertical equatorial section of dissolved oxygen of CCSM3 for the PTB for the (a) 4 x CO₂ experiment, (b) 12.7 x CO₂, and (c) 12.7 x CO₂ WC experiment. Contour interval is 0.5 ml L⁻¹. Note that the left ocean basin is the Tethys Ocean and right ocean basin denotes the Panthalassa. Hypoxic conditions are light gray shaded and anoxic conditions are dark gray shaded.

Fig. 10. Flowchart model of cause-and-effect relationships during the PTB extinction crisis and Early Triassic recovery intervals. Modified from Wignall (2001) and Algeo et al. (2011) cloud optical processes discussed in this have been added to the graph. Note that according to Berner (2002) a short residence time of 12±3 years (Schimel et al., 2005) for methane in the atmosphere is assumed and that methane is eventually oxidized to CO₂ and H₂O. Different time scales for the various processes apply. For example, processes related to cloud optics are in time scale of hours to a few years whereas weathering are in time scale of 10,000 to million of years.