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Response of Central European SST to atmospheric pCO_2 forcing during the Oligocene – A combined proxy data and numerical climate model approach

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CO_2-induced global warming will affect seasonal to decadal temperature patterns. Expected changes will be particularly strong in extratropical regions where temperatures will increase at faster rates than at lower latitudes. Despite that, it is still poorly constrained how precisely short-term climate dynamics will change in a generally warmer world, particularly in nearshore surface waters in the extratropics, i.e., the ecologically most productive regions of the ocean on which many human societies depend. Specifically, a detailed knowledge of the relationship between pCO_2 and seasonal SST is crucial to understand interactions between the ocean and the atmosphere. In the present investigation, we have studied for the first time how rising atmospheric pCO_2 levels forced surface temperature changes in Central Europe (paleolatitude ~45 °N) during the mid-Oligocene (from ca. 31 to 25 Ma), a time interval of Earth history during which global conditions were comparable to those predicted for the next few centuries. For this purpose, we computed numerical climate models for the Oligocene (winter, summer, annual average) assuming an atmospheric carbon dioxide rise from 400 to 560 ppm (current level to two times pre-industrial levels, PAL) and from 400 to 840 ppm (= three times PAL), respectively. These models were compared to seasonally resolved sea surface temperatures (SST) reconstructed from $\delta^{18}O$ values of fossil bivalve shells (*Glycymeris planicostalis*, *G. obovata*, *Palliolum pictum*, *Arctica islandica* and *Isognomon maxillata sandbergeri*) and shark teeth (*Carcharias cuspidata*, *C. acutissima* and *Physogaleus latus*) collected from the shallow water deposits of the Mainz and Kassel Basins (Germany). Multi-taxon oxygen isotope-based reconstructions suggest a gradual rise of temperatures in surface waters (upper 30 to 40 m), on average, by as much as 4 °C during the Rupelian stage followed by a 4 °C cooling during the Chattian stage. Seasonal temperature amplitudes increased by ca. 2 °C during the warmest time interval of the Rupelian stage, with warming being more pronounced during summer (5 °C) than during winter (3 °C). According to numerical climate simulations, the warming of surface waters during the early Oligocene required a CO_2 increase by at least 160 ppm, i.e., 400 ppm to 560 ppm. Given that atmospheric carbon dioxide levels predicted for the near future will likely exceed this value significantly, the Early Oligocene warming gives a hint of the possible future climate in Central Europe under elevated CO_2 levels.

1. Introduction

Unabated CO_2 emissions will result in atmospheric concentrations of this greenhouse gas of over 900 ppm by the end of the current century (Collins et al., 2013) and up to 1800 ppm 300 years later (Zachos et al., 2008). In this scenario, surface temperatures are expected to increase significantly, specifically at higher latitudes, causing polar ice to melt and eventually disappear (e.g., Körper et al., 2013). Reduced meridional temperature gradients will lead to substantially different circulation patterns (e.g., Cai and Chu, 1998; Hansen et al., 2004) with major repercussions on seasonal to multi-annual climate dynamics (e.g., Marshall et al., 2001; Solomon et al., 2007), i.e., within the time-scales of human perception.

Despite the relevance for human societies and the global economy, it is still not well constrained how short-term climate variability will change in a generally warmer world (e.g., Collins et al., 2010). For example, will summers become hotter and winters colder, and how much for a given atmospheric pCO_2? How will seasonal temperature amplitudes change in nearshore surface waters? These settings are the ecologically most productive regions of the ocean on which many human societies depend, and they are particularly sensitive to short-term climate

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fluctuations, because of their low water-mass inertia and shallow water depth. Such questions can be addressed by studying past climates during which boundary conditions were broadly similar to those today and predicted for the near future. For this purpose, the Oligocene epoch serves as a suitable candidate. The distribution of land and sea was comparable to the present-day situation (e.g., Scotese et al., 1988; Rögl, 1999). During the early Oligocene, CO_2 levels and surface temperatures were slightly higher than in modern times (Henderiks and Pagani, 2008; Cramer et al., 2011). Although this time interval could help us to obtain a better perspective on future, anthropogenically forced climates, our knowledge on seasonality during the Oligocene and its relationship with pCO_2 is extremely limited.

Atmospheric pCO_2 levels of the Oligocene have been inferred from $\delta^{13}C$ values of sedimentary biomarkers such as algal alkenones (e.g., Pagani et al., 2005; Henderiks and Pagani, 2008; Zhang et al., 2013), boron isotopes of foraminiferan tests (e.g., Pearson et al., 2009; Raitzsch and Honisch, 2013) and stomatal density of fossil leaves (e.g., Retallack, 2001; Roth-Nebelsick et al., 2012). Based on data recorded in these archives, atmospheric pCO_2 oscillated between ca. 200 and ca. 1000 ppm. However, these reconstructions come with large uncertainties (Beerling and Royer, 2011). Likewise, numerical modeling does not provide more precise estimates and suggests that atmospheric pCO_2 may have fluctuated between ca. 300 ppm and ca. 800 ppm, i.e., the threshold of the northern and southern hemisphere polar glaciations, respectively (DeConto and Pollard, 2003; Pollard and DeConto, 2005; DeConto et al., 2008).

The great majority of existing Oligocene temperature reconstructions are based on geochemical records preserved in biogenic archives of the deep sea. For example, the $\delta^{18}O$ signature of deep-sea benthic foraminifera ($\delta^{18}O_{BF}$) has been used to track long-term climate trends (e.g. Miller et al., 1991; Zachos et al., 2001; Pekar et al., 2002; Wade and Pälike, 2004; Pälike et al., 2006). Information on how temperature fluctuated on shorter time-scales is largely limited to air temperatures. According to Pross et al. (1998, 2001), seasonal air temperatures in Central Europe during the late Rupelian oscillated between 7 and 10 °C in winter, and 26 and 28 °C in summer. These values agree well with the air temperature data (5 to 26 °C; average: 16–17 °C) provided by Mosbrugger et al. (2005) and Erdei et al. (2012). However, seasonally resolved temperature data from surface ocean waters is virtually absent. The $\delta^{18}O$ values of shark teeth (Tütken, 2003) and foraminiferan tests (Grimm, 1994) yield temperature ranges of 12 to 22 °C in surface waters, and 6 to 15 °C in settings below the thermocline (ca. 150 m; Grimm et al., 2011). More recently, Walliser et al. (2015) demonstrated for the first time that $\delta^{18}O$ values of well-preserved bivalve shells from the Rhenish triple junction (Central European epicontinental seaway) can provide subseasonally resolved records of temperature that prevailed in surface waters at mid-latitudes during the Early Oligocene. Up until now, has not been resolved how seasonal temperatures in surface waters evolved during the Oligocene.

Inspired by the proof-of-concept study by Walliser et al. (2015), we investigated how seasonal sea surface temperatures (SST) in central Europe changed during selected time slices of the Rupelian and Chattian stages. Reconstructions were based on $\delta^{18}O$ values of well-preserved shells of five different bivalve species (*Glycymeris planicostalis, G. obovata*, *Arctica islandica*, *Isognomon maxillata sandbergeri* and *Palliolum pictum*), together with information on their growth patterns, as well as shark teeth. Fossils came from shallow marine sediments of the Rhenish triple junction, i.e., the Mainz and Kassel Basins. Furthermore, numerical climate models were computed to estimate how winter, summer and average Oligocene SST change when atmospheric pCO_2 increases from 400 ppm to 560 ppm (=2× PAL, preindustrial atmospheric level) and 840 ppm (3× PAL), respectively. These data were then compared to proxy-based SST reconstructions in order to further constrain Oligocene atmospheric pCO_2 levels. A detailed knowledge of the relationship between pCO_2 and seasonal SST is crucial to understand interactions between the ocean and the atmosphere.

2. Geological setting, material and methods

2.1. Geological setting

The Mainz Basin and the Kassel Basin are located at the northwestern margin of the Upper Rhine Graben and the central part of the Hesse Depression (Fig. 1A), respectively. Their formation is associated with the taphrogenesis of the European Cenozoic Rift System (Sissingh, 2003; Dèzes et al., 2004). Due to their geographical vicinity and the absence of major tectonic boundaries between them, the two basins experienced similar sedimentary histories which encompass marine, brackish and terrestrial deposits covering a time interval of ca. 20 Ma (Lower Eocene to Early Miocene; Grimm et al., 2011; Ritzkowski et al., 2011). During the Oligocene, repeated marine transgressions occurred from the north (Berger et al., 2005a, 2005b) which were caused by local subsidence and eustatic sea level rise (Fig. 1B–D). As a consequence, the Mainz Basin and Kassel Basin became part of an intracontinental marine strait that extended from the paleo-North Sea to the Alpine Belt (Picot, 2002; Berger et al., 2005a) (Fig. 1B). The largest marine ingression took place during the upper Rupelian stage (Fig. 1C), i.e., between sea-level high stand Ru2/Ru3 (~32 Ma) and Ru3/Ru4 (~29.5 Ma) of Haq et al. (1988). It possibly resulted in a (temporary) connection of the southern Upper Rhine Graben (URG) system with the western Molasse Basin (Martini, 1982; Picot, 2002; Berger et al., 2005a, 2005b; Grimm, 2006). In the Mainz Basin, this marine phase is represented by nearshore, coarse-grained siliciclastic deposits of the Alzey Formation and coeval basinal pelites of the Bodenheim Formation as well as the overlying clays and fine sands of the Stadecken Formation (Grimm et al., 2000; Berger et al., 2005b; Grimm et al., 2011). The studied bivalve shells originate from sediments of the Alzey Formation and the Stadecken Formation (Albig Bank) (Figs. 1D and 2).

Age control of the nearshore strata of the Alzey Formation is mainly based on the correlation with basinal pelites of the Bodenheim Formation (Grimm et al., 2000). Age control of the latter is largely based on calcareous nanoplankton (Martini and Müller, 1971; Grimm et al., 2011), dinoflagellate cysts (Pross, 2001; Grimm et al., 2011) and to a lesser extent, benthic foraminifera (Grimm, 1998, 2002) (Fig. 2). With regard to the nanoplankton zonation, the Alzey Formation comprises the upper part of NP23 and the lower part of NP24 (Grimm, 1994; Berger et al., 2005b). Furthermore, the presence of the benthic foraminiferan, *Turrilina alsatica* in the lowermost deposits of the Bodenheim Formation (Grimm et al., 2000, 2011) places the onset of fully marine sedimentation in the Mainz Basin within the benthic foraminifera zone NSR 7b (Hardenbol et al., 1998) (Fig. 2). Thus, the lower boundary of the Alzey Formation falls about in the middle of the nanoplankton zone NP23 (Grimm et al., 2011). Based on benthic foraminiferan associations, the Alzey Formation can be subdivided into a lower *Planorbulina difformis–Cibicides lobatus* Abundance Zone (here referred to as lower Alzey Formation) belonging to NP23, and an upper Miliolid Abundance Zone (here referred to as upper Alzey Formation) which comprises the upper part of NP23 and the lower part of NP24 (Grimm, 1998; Grimm, 2002). Numerical $^{87}Sr/^{86}Sr$ dating performed on a well-preserved bivalve shell collected from the Miliolid Zone yielded an age of 30.1 ± 0.1 Ma (Grimm et al., 2003).

The overlying deposits of the Stadecken Formation belong exclusively to the nanoplankton zone NP24 (Martini and Müller, 1971; Grimm et al., 2011) and are capped by a wide-spread tempestite horizon (the so-called Albig Bank) which comprises the lower part of the *Chara microcera* zone (Grimm et al., 2000). The upper boundary of the Stadecken Formation is delimited by the onset of the brackish facies of the overlying Sulzheim Formation (Grimm et al., 2011) which comprises in its lowermost part the fossil mammal assemblages belonging to the Paleogene mammal zone MP24 (Bahlo, 1976; Bahlo and Neuffer, 1978; Bahlo and Tobien, 1982). Accordingly, the deposition of the Albig Bank probably occurred