Extinquishing a Permian World

Elke Schneebeli-Hermann
Palaeoecology, Department of Physical Geography, Utrecht University, Budapelaan 4, 3584 CD Utrecht, Netherlands

At the end of the Permian, ca. 252 Ma ago, marine and terrestrial fauna were facing the most extensive mass extinction in Earth history (Raup and Sepkoski, 1982). 80%–95% of all species on Earth, on land and in the oceans, became extinct (Benton et al., 2004) within an estimated time interval of less than 200 k.y. to 700 k.y. (Huang et al., 2011; Shen et al., 2011). Among the prominent Paleozoic animal groups that vanished are fusulinid foraminifera, rugose and tabulate corals, and the arthropod class Trilobita. The numerous hypotheses about the causes of the mass extinction include various environmental changes, mostly related to the emplacement of the Siberian Traps large igneous province. The compilation of radiometric U/Pb ages for the mass extinction and the Siberian Traps demonstrate a temporal overlap of both events (Svensen et al., 2009). A prominent hypothesis for the mass extinction is an accentuated global climate change scenario induced by volcanic CO2 degassing (e.g., Svensen et al., 2009) that triggered biotic responses in the sea and on land. But what do we know about the climate at this time in Earth history?

The two main features of climate recorded in the geological archives are temperature and humidity i.e., the moisture that is available for plant growth, and to a certain extent rainfall patterns. Climate simulations model the general circulation patterns during Permian–Triassic times. The paleogeography was characterized by the supercontinent Pangaea extending nearly from pole to pole, with large land masses in the mid-latitudes of the Northern and Southern Hemispheres, and with the Tethys Ocean in the tropics (e.g., Smith et al., 1994). Climate models and sensitivity experiments demonstrated that this paleogeography provided the preconditions for a monsoonal circulation with strong seasonality of temperatures and rainfall on the Tethyan coasts, and distinct northern and southern intertropical convergence zones (Fig. 1) (e.g., Kutzbach and Ziegler, 1993; Parrish, 1993). Moist conditions prevailed in middle and high latitudes and along the western Pangaea coast, contrasting with the year-round arid Pangean tropics. (Kutzbach and Ziegler, 1993; Parrish, 1993). Recent climate modeling for the Permian–Triassic scenario tested the impact of rapid temperature change and showed that shifts in biomes were less pronounced during global warming compared to global cooling (Roscher et al., 2011). For the Late Permian oceans, sensitivity experiments with coupled atmosphere-ocean models imply that elevated CO2 levels caused significant temperature increases in high-latitude sea-surface water masses, leading to reduced global sea-surface temperature (SST) gradients (e.g., Kiehl and Shields, 2005).

A proxy record for ancient SSTs is stable oxygen isotope data of marine biomineralizing fauna. During biomineralization, marine fauna incorporate dissolved oxygen into their hard parts (e.g., calcite, aragonite, apatite) in or close to isotopic equilibrium with the ambient sea water (Urey, 1947; Epstein et al., 1951). Therefore, geochemical signals preserved in marine invertebrate shells and skeletal parts can be used as a proxy for SST, if the isotope composition of the ambient water is known (although isotope fractionation during biomineralization and diagenetic alteration need to be considered). Oxygen isotopes measured on unaltered brachiopod shells from the Late Permian Bellerophon Formation, and from a single sample from the overlaying basal Werfen Formation, in southern Italy were interpreted to represent a rise of 6–10 °C of tropical SST (Kearsey et al., 2009 and references therein). However, the resolution and stratigraphic constraints of these data are limited. In the search for unaltered archives for oxygen isotope records for paleotemperature reconstructions, the significance of biogenic apatite of tooth enamel is increasing. The oxygen isotope signature of conodont apatite is little affected by post-depositional changes (Joachimski and Buggisch, 2002). New analytical techniques and better understanding of isotopic changes during diagenesis have revealed the importance of the conodont paleothermometry for the reconstruction of ancient ocean temperatures (e.g., Barham et al., 2012; Vennemann et al., 2002).

The work of Joachimski et al. (2012, p. 195 in this issue of Geology) uses the conodont paleothermometry to reconstruct ocean temperatures of a tropical site during the largest mass extinction in Earth’s history. One of the best studied Permian–Triassic successions is the Global Boundary Stratotype Section and Point (GSSP) in Meishan, South China. Joachimski et al. measured the oxygen isotope values of P1 elements of the conodont genera Clarkina (or Neogondolella) and Hindeodus recovered from the Meishan and Shangsi sections, choosing a δ18O value of −1‰ Vienna standard mean ocean water for their temperature calculation. They demonstrate that in the Meishan area, SST increased by 1–5 °C, from an average temperature of 22 °C, before the main extinction event. Thereafter, temperatures remained fairly stable up to the extinction event, where oxygen isotopes indicate another distinct warming by 5–8 °C, up to SST of 32–35 °C in the earliest Triassic. The data of Joachimski et al. are an important advance in the discussion of the Permian–Triassic climate change, as they quantify SST change in a stratigraphically well-constrained framework. The documented oxygen isotope shift (i.e., the warming) coincides with a negative shift in carbon isotope ratios over a period of possibly only ~110 k.y. (Shen et al., 2011).
Permian macroflora (distribution of biomes) largely support the general climate zone distribution with strong seasonality for the Late Permian, as derived from climate models (Rees et al., 2002). The globally less-differentiated Early Triassic flora were interpreted to indicate warm-temperate climate up to 70°N (Ziegler et al., 1993). Unfortunately, the time constraint of most macrofloral records to infer detailed climate evolution is limited by the lack of calibration. Higher chronological resolution is achieved by implementing palynological data. Despite uncertainties regarding the botanical affinities of 250 Ma spores and pollen, the approach of distinguishing sporomorphs according to the requirements of their parent plants with respect to water availability has been applied successfully to describe relative changes in humidity (e.g., Hochuli et al., 2010). In Norway, chem stratigraphically calibrated palynological records indicate a distinct shift from pollen-dominated (conifers and seed ferns) to spore-dominated (ferns and lycophytes) assemblages across the Permian–Triassic transition. This change has been interpreted as a shift to more humid conditions in the earliest Triassic (Hochuli et al., 2010). Also, in the monsoon-dominated region of the southern subtropics, increasing spore abundances occur toward the Early Triassic (Hermann et al., 2012). Thus, the warming in the latest Permian is associated with changing global and regional precipitation and evaporation patterns resulting in more humid conditions in the northern mid-latitudes and southern subtropics.

The distribution and diversity of ammonoids in space and time, introduced as the latitudinal gradient of generic richness (LGG), have also been interpreted to reflect latitudinal SST gradients (Brayard et al., 2006). For the earliest Triassic (Griesbachian), the LGG is very low, which has initially been interpreted as reflecting a warm equable global climate corresponding to flat SST gradients. This supports the climate models predicting a lower SST gradient with increased CO₂ levels. However, in the earliest Triassic, this approach is limited because the diversity of ammonoid communities was also affected by the extinction event.

Evidence for the modeled monsoon circulation during Permian–Triassic times has been inferred from sedimentary records (e.g., Mutti and Weissert, 1995). Widespread redbed deposits are assumed to represent areas with seasonal rainfall (e.g., Parrish, 1993). For the Late Permian, a zone of high evaporation in the tropics can be delineated by mapping the global distribution of evaporites, reef carbonate, and coal deposits (Ziegler et al. 2003) (Fig. 1). Correlation of terrestrial sequences with marine successions is hampered by the different biostratigraphic frameworks, but nevertheless the change from paleosols with coals to paleosols containing green–red-mottled claystones in Antarctica has been interpreted as a change from a dryer, cooler climate in the Late Permian to more humid, warmer conditions in the Early Triassic (Retallack and Krull, 1999).

In summary, the data of Joachimski et al. show that in the tropics, SST started to increase in the latest Permian prior to the main extinction event, continued increasing during the extinction into the earliest Triassic, and may have reached values of >32 °C. Additional paleotemperature records of comparable quality are now needed from less-condensed successions (i.e., with higher stratigraphic resolution) to enable us to further disentangle biotic and abiotic events around the Permian–Triassic boundary. Furthermore, paleotemperature data from other proxies may in turn strengthen the use of conodont paleothermometry in Permian to Triassic climate reconstructions.