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Effect of aridification on carbon isotopic variation and ecologic evolution at 5.3 Ma in the Asian interior

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ABSTRACT

The Cenozoic era is marked by dramatic climatic and ecological changes. The timing of the emergence and the subsequent expansions of C_4 grasses are prominent biological events on Earth. In China, thick Cenozoic deposits in the Tarim and Junggar Basins, which are located in the Asian interior, provide important geological archives for studying paleoenvironmental changes. Here we use carbon isotope compositions of organic matter to reconstruct the history of ecologic evolution during the late Cenozoic in the Tarim and Junggar Basins. The results show that there is a shift to slightly higher $\delta^{13}C$ values at 5.3 Ma indicating a change in terrestrial ecosystems in the Asian interior driven by an increased regional aridity rather than decreasing atmospheric pCO_2 levels. The weakened water vapor transportation related to the retreat of Paratethys Ocean and the enhanced rain shadow effect of mountain uplift during the latest Miocene mostly triggered this event.

1. Introduction

The Cenozoic world climate has prominent stepwise cooling and drying trends, marked by the initiation of ice sheet in the Antarctic since the Eocene–Oligocene boundary at 34 Ma (e.g., Kennett, 1977; Miller et al., 1991; Zachos et al., 2001) and the emergence of high latitude sea ice in the northern hemisphere in the late Cenozoic at about 7–5 Ma (e.g., Jansen and Sjøholm, 1991; Larsen et al., 1994). Over the same time span, grasslands expanded in terrestrial ecosystems (Jacobs et al., 1999). Although the emergence of C_4 plants on earth may be as early as 32 to 23 Ma (e.g., Fox and Koch, 2003; Osborne and Beerling, 2006; Urban et al., 2010; Edwards et al., 2010), the significant C_4 expansion occurred much later, mostly from the latest Miocene to Pliocene or even into the Pleistocene (e.g., Quade et al., 1989; Cerling et al., 1993, 1997; Dettman et al., 2001; Retallack, 2001; Bywater-Reyes et al., 2010; Fox et al., 2012a, 2012b).

However, there have been different opinions about the causes of C_4 expansion in the late Neogene. Quade et al. (1989) linked the dramatic ecologic shift in the latest Miocene in Pakistan to a marked strengthening of the Asian monsoon system. Cerling et al. (1997) proposed that a drop in atmospheric CO_2 levels drove this expansion, but the later established CO_2 record indicates more or less steady levels throughout the Neogene (e.g., Pagani et al., 2005). Pagani et al. (1999) attributed the C_4 plant expansion to enhanced regional aridity or changes in seasonal precipitation patterns. Other alternative views include that grasslands were a biological force in their own right (Retallack, 1998) or the worldwide rapid increase in C_4 ecosystem responded to both climate and tectonics (Kohn and Fremd, 2008).

Not only the causes for the late Miocene C_4 expansions are controversial, but also the exact timing of this ecological event is still debated. Quade et al. (1989) reported dramatic ecological shift beginning ~7.4–7.0 Ma in Pakistan. Cerling et al. (1993) concluded that C_4 expansions occurred between 7 and 5 Ma in Pakistan and North America, and later Cerling et al. (1997) suggested a global scale C_4 expansion at 8–6 Ma. Latorre et al. (1997) suggested that the presence of C_4 plants started at 7.3–6.7 Ma in Argentina. Fox and Koch (2004) indicated that the percentage of C_4 grasses in the Great Plains of the United States increased from 6.4 to 4.0 Ma; later, more records in the central Great Plains indicate C_4 expansion from the early late Miocene or Pliocene to the early Pleistocene (e.g., Martin et al., 2008; Fox et al., 2012a, 2012b). Additionally, $\delta^{13}C$ values of n-alkanes from the Gulf of Mexico indicate that terrigenous C_4 plants steadily increased during the late Miocene into the Pleistocene (Tipple and Pagani, 2010). The recent higher-resolution study on the well-known Siwalik Group of Pakistan revealed a more gradual transition between 8.0 and 4.5 Ma in which C_3 and C_4 plants occupied different subenvironments of the Siwalik alluvial plain (Behrensmeyer et al., 2007). In South Africa, C_4 grasses became a significant part of the Makapansgat Valley ecosystem at approximately 4–5 Ma (Hopley et al.,

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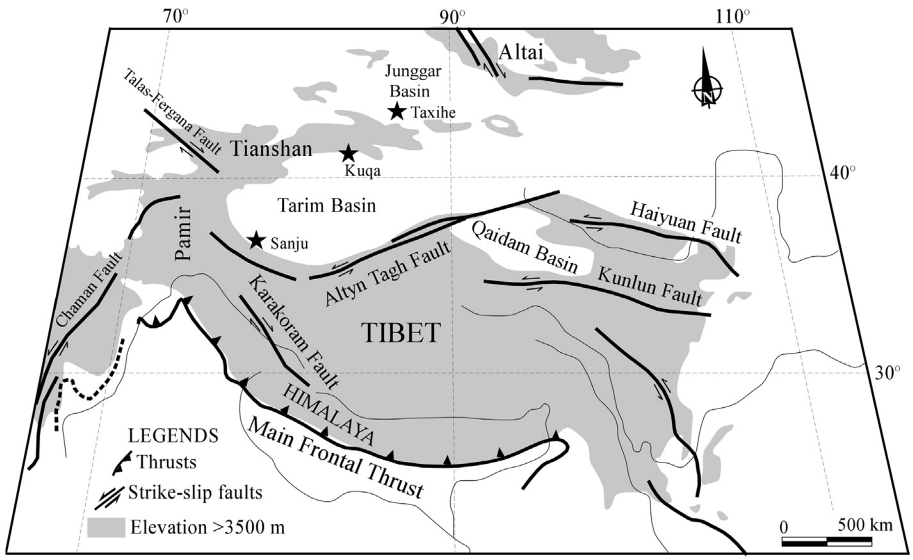

Fig. 1. Simplified geological map shows the main active faults of central Asia as well as the locations (black stars) of the sections mentioned in the text (modified from Avouac and Tapponnier, 1993).

2007). In central North China, Ding and Yang (2000) defined a major expansion of C_4 plants at ca. 4.0 Ma. Wang and Deng (2005) and Biasatti et al. (2010) presented extensive isotope data and suggested that C_4 grasses were either absent or insignificant in the Linxia Basin prior to ∼2–3 Ma and only became a significant component of local ecosystems in the Quaternary. Significant C_4 biomass was also reported in the Gyirong Basin in southern Tibet at ∼7 Ma (Wang et al., 2006) and in the central Inner Mongolia from ∼8 to 4 Ma (Zhang et al., 2009). In South China Sea, the $\delta^{13}C$ record of black carbon of marine sediments indicates an increasing trend since 8.7 Ma (Jia et al., 2003).

The C_4 expansion can be studied by using carbon isotopic compositions of sediments. Previous studies have demonstrated that the C_3 and C_4 photosynthetic pathways fractionate carbon isotopes to different degrees; C_3 and C_4 plants have averaged $\delta^{13}C$ values of −27‰ and −13‰, respectively (Cerling et al., 1997). Different from the soil carbonate and mammalian fossil tooth enamel which are significantly enriched in $\delta^{13}C$ compared to source carbon, organic matter preserves the isotopic distinction with little or no isotopic fractionation (Cerling et al., 1989).

In China, huge inland basins occupy northwestern China. The remote distances to oceans (both the Pacific and the Atlantic oceans) make this region the driest region in the interior of Asia. To date, there are only limited carbon isotope records available from this region (e.g., Wang and Deng, 2005; Charreau et al., 2012; Zhang et al., 2009, 2012). In this paper, we report the results of stable carbon isotope analyses of organic matter in order to discuss the timing and driving mechanisms of significant late Cenozoic ecological changes in the interior of Asia.

2. Geological setting

The Asian interior consists of several east–west trending mountain ranges (e.g., Kunlun Mountains, Tianshan Mountains), forming a series of mountain-basin systems. In this paper, we focus on studying the past ecological changes in the Tarim and Junggar Basins (Fig. 1). The Tarim Basin is the largest inland basin in China and is constrained by three large mountain ranges: the Tianshan range to the north, the Pamir Mountain to the west, and the Kunlun range to the south (Fig. 1). Located in the rain shadow of the Tibetan Plateau, the climate is extremely dry with annual rainfall of less than 50 mm in the center of the basin, and it is the driest region in the interior of Asia. The main part of the basin is occupied by the Taklimakan Desert, which covers an area of 132,000 square miles (342,000 km^2) and is the world second largest shifting sand desert on Earth (Zhu et al., 1980).

The Junggar Basin is constrained by the Altay Mountains to the northeast and the Tianshan Mountains to the south (Fig. 1). This basin has an area of 380,000 km^2 and the center of the basin is the Gurbantunggut Desert. The annual precipitation ranges from 70 to 250 mm and it is a steppe and semi-desert basin.

Both basins are of structural origins. The Cenozoic uplifts of the surrounding mountains are related to the continuing convergence of India and Eurasia since their collision in the Eocene (Molnar and Tapponnier, 1975). Being coupled with the mountain uplift, both basins experienced major subsidence during the Cenozoic era. The Cenozoic sedimentary rocks derived from the surrounding mountain belts accumulated in the foreland basins with a thickness of up to 10 km (e.g., Li et al., 1996; Jia, 1997). The tectonically deformed Cenozoic sediments not only provide useful constraints on the mountain building in the surrounding orogens but also serve as an important archive for studying past ecological evolution.

In this study, two sections were chosen for studying paleovegetation changes. Among them, the Sanju (37°11′N, 78°29′E) section lies in the southern edge of the Tarim Basin (Fig. 1), while the Taxihe section (44°06′N, 86°20′E) is located in the southern margin of the Junggar Basin (Fig. 1).

The strata of the above sections are all tectonically deformed by thrusting and folding in the foreland basins (Fig. 2). The studied deposits at Sanju span an age range from late Miocene to Quaternary (Sun and Liu, 2006), with a thickness of 1626 m, and consist of the late Miocene Pakabulake Formation, the Pliocene Artux Formation, and the latest Pliocene to early Pleistocene Xiyu Formation (Fig. 2a).

The Taxihe section is an overturned anticline, with steep or overturned strata in the core of the northern limb and a gently dipping southern limb (Fig. 2b), consisting of the Oligocene Anjihaihe Formation (E3a), the latest Oligocene Shawan Formation (E3s), the Miocene Taxihe Formation (N1t), the Pliocene Dushanzi Formation (N2d), and the early Pleistocene Xiyu (Q1x) Formation (Fig. 2b). The Taxihe section is 2960 m thick (Sun and Zhang, 2009). Generally, there is an up section trend towards coarser grain size from the late Oligocene Shawan to the early Pleistocene Xiyu Formations.