

Research Paper

Early Jurassic carbon cycle perturbations recorded in terrestrial sediments of the Sichuan Basin, China

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ABSTRACT

The Toarcian Oceanic Anoxic Event (T-OAE, ~183 Ma) is marked in the sedimentary record by a sharp negative carbon isotope excursion, thought to be causally linked to the emplacement of the Karoo-Ferrar Large Igneous Province and the associated release of ^{12}C -enriched carbon. The T-OAE coincided with global climate and environmental changes, as well as biotic events, indicating substantial modifications in ecosystems. Recent studies have focused on the evidence of geological responses to the T-OAE in Early Jurassic terrestrial basins in China, particularly the Sichuan Basin. Nevertheless, debate remains over the identification of this event, owing to inadequate age-constraints of many sections, and a lack of robust correlations of the carbon isotope records. Moreover, the long-term evolution of the terrestrial carbon isotope record through the Early Jurassic, and its correlation to marine records, is still not firmly established. In this paper, we present new carbon isotope analyses of carbonate ($\delta^{13}\text{C}_{\text{carb}}$) from lacustrine carbonates and terrestrial organic matter ($\delta^{13}\text{C}_{\text{org}}$) from bulk rocks within the Ma'anshan and Da'anzhai members of the Ziliujing Formation from the Dacao 'D' (DCD) section in the eastern Sichuan Basin. Palynological-palynofacies analysis reveals a predominance of *Classopollis* pollen together with marker taxa such as *Ischyosporites variegatus*, *Contignisporites problematicus*, in the palynological assemblage, indicating a Pliensbachian–Toarcian age. A negative carbon isotope excursion (NCIE) is recorded in the organic carbon isotope data at the topmost part of the Pliensbachian Ma'anshan Member, which can be correlated to the Pliensbachian–Toarcian Boundary Event. This is followed, in the Toarcian Da'anzhai Member, by a major NCIE recorded in both organic matter and carbonate carbon isotope data which can be correlated to the T-OAE NCIE. A long-term carbon isotope record spanning the Sinemurian to Toarcian in Sichuan terrestrial sediments is also been reconstructed and its correlation with coeval marine records is proposed. A broader review of $\delta^{13}\text{C}$ data from Chinese terrestrial basins spanning the Pliensbachian–Toarcian highlights a distinct ^{13}C -depleted signature in the Sichuan Basin compared to basins at higher latitudes. Changes in latitudinal gradients and organic matters in the lake sediments were likely important factors influencing the amplitudes of the T-OAE NCIE and the carbon isotope values in terrestrial sedimentary records.

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1. Introduction

The Early Jurassic Toarcian Oceanic Anoxic Event (T-OAE; also named the Jenkyns Event *sensu* Reolid et al., 2020) was first identified on the basis of the widespread occurrence of black shale deposits on the European continental shelf, which were inferred

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to have formed within anoxic water (Jenkyns, 1985, 2010). After decades of study, the T-OAE has been further recognized as a major hyperthermal event, when seawater temperature could have risen by up to 5–10 °C (e.g., Bailey et al., 2003; Ruebsam et al., 2020a). This warming was accompanied by a crisis in shallow water carbonate depositional systems (Trecalli et al., 2012; Krencker et al., 2020), and enhanced weathering and biotic turnover on both land and in the oceans (e.g., Percival et al., 2016; Slater et al., 2019; Kemp et al., 2020; Pol et al., 2020; Reolid and Ainsworth, 2022; Reolid et al., 2022; Baranyi et al., 2023, 2024). The T-OAE is marked in the stratigraphic record by a major perturbation of the global carbon cycle, manifested as a negative carbon isotope excursion (NCIE) in terrestrial and marine sedimentary records, including marine carbonates, and marine and terrestrial organic carbon (e.g., Kemp et al., 2005; Hesselbo et al., 2007; Xu et al., 2017; Ruebsam et al., 2020b; Jin et al., 2020, and reference therein). The T-OAE NCIE has been attributed to the release of a large amount of ^{12}C -enriched carbon into the ocean–atmosphere system, perhaps originating from: (1) thermogenic methane released due to the intrusion of magmas into organic-rich rocks during the emplacement of the Karoo and/or Ferrar Large Igneous Province (e.g., McElwain et al., 2005; Svensen et al., 2007; Kemp et al., 2024), (2) the massive release of methane from gas hydrate of marine and terrestrial sediments (e.g., Hesselbo et al., 2000; Kemp et al., 2005; Them et al., 2017), and/or (3) the melting of carbon-rich cryosphere reservoirs (e.g., Ruebsam et al., 2020b).

The T-OAE had a significant impact both on marine and terrestrial systems (e.g., Slater et al., 2019; Reolid et al., 2022; Baranyi et al., 2023, 2024, and references therein). As such, exploring the nature of the T-OAE in terrestrial settings can provide valuable insights into the way terrestrial ecosystems and environments responded to large-scale carbon release and warming, thus providing potential insights into the environmental responses expected at the present day under continued anthropogenic carbon emissions. Importantly, however, when compared to the wealth of data on the T-OAE from marine records, our knowledge of the T-OAE from terrestrial sedimentary records is still limited.

During the Jurassic, many mega-lake systems existed in South Eastern Eurasia (Fig. 1A). Extensive sedimentary record from these lacustrine systems gives the opportunity to study the T-OAE and its

impact on the terrestrial environment. In particular, Toarcian lacustrine deposits in the Sichuan Basin, which lied within the boreo-tropical climate zone, are well-preserved (e.g., Li et al., 2020). The depositional age of the Da'anzhai Member in this basin (the fourth member of Ziliujing Formation) has been constrained as Toarcian based on Re-Os isotope dating (180 ± 3.2 Ma) and palynology, e.g., a superabundance of the pollen *Classopollis* sp. and the occurrence of *Halosphaeropsis liassica* (Xu et al., 2017). However, Liu et al. (2020b) argued that the age determination of the T-OAE in the Sichuan Basin, as proposed by Xu et al. (2017), lacks constraint owing to the absence of unambiguous age-diagnostic palynomorph taxa. Instead, Liu et al. (2020b) suggested that the majority of the intervals studied by Xu et al. (2017) were deposited after the T-OAE, e.g., in the late early to middle-late Toarcian. Although the age determination of the inferred T-OAE event in the Sichuan Basin remains debated (e.g., Liu et al., 2020b), there is clear evidence that the organic-rich mudstones/shales of the Da'anzhai Member record an abrupt NCIE, with magnitudes of ca. -4‰ to -6‰ (Xu et al., 2017; Liu et al., 2020a).

Primarily, the debate surrounding the age of the NCIE in the Da'anzhai Member stems from the lack of robust chronostratigraphic and biostratigraphic constraints. At the same time, however, the lack of a comprehensive, long-term carbon isotope record through the Toarcian strata in the basin further limits our knowledge of the overall context of the NCIE, and the long-term paleoenvironmental evolution of the basin. In addition, fossiliferous limestones form one of major lithological components of the Da'anzhai Member, but these have not been investigated in detail. This is a key knowledge gap because lacustrine biogenic and abiogenic carbonates have been demonstrated to be useful materials for reconstructing the carbon cycle during major geological events, e.g., Paleocene–Eocene Thermal Maximum (Chen et al., 2014).

Here we present a high-resolution carbon isotope record from mudstones/shales and lacustrine carbonates of the Da'anzhai Member in the Sichuan Basin. We couple this with a detailed palynofacies and palynostatigraphical characterization of the Da'anzhai Member to constrain the age of this new record. In addition, carbon isotope studies spanning the Pliensbachian–Toarcian from different terrestrial basins at different latitudes in China have been reviewed. Finally, a long-term carbon isotope-based correlation

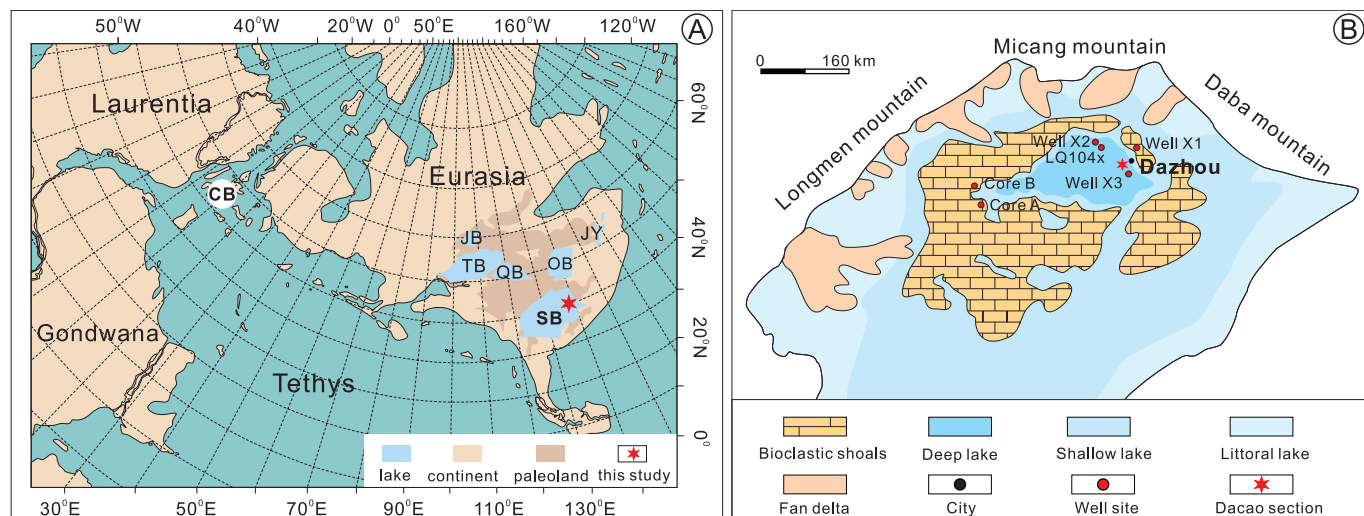


Fig. 1. Location maps of the study area. (A) Early Jurassic paleogeographic map showing the major Chinese terrestrial basins, including Jinyang (JY), Tarim (TB), Qaidam (QB), Ordos (OB), Sichuan (SB) basins, as well as the Cleveland Basin (CB) in the UK. Please note that the paleogeographical position and extent of these basins is an approximation. The map is modified from Xu et al. (2017). (B) The paleogeographic map of the Sichuan Basin illustrates the facies associations and surrounding mountain ranges during the deposition of the Da'anzhai Member. The locations of Dazhou city, Dacao section, and adjacent cores, e.g., Cores A and B of Xu et al. (2017); Well X1, X2, X3 of Cui et al. (2023); LQ104X of Liu et al. (2020), are marked in the map. The map is modified from Cui et al. (2023).

between terrestrial and marine sedimentary sequences spanning the Sinemurian to Toarcian is proposed.

2. Geological setting

The Sichuan Basin is located on the western margin of the South China Block and transitioned from a Paleozoic–Middle Triassic passive continental margin to a continental clastic sedimentary basin in the Late Triassic, due to the collision between the South China and North China blocks along the Mianlue suture zone (e.g., Li et al., 2014; Liu et al., 2021). During the Early Jurassic, the Sichuan Basin formed a mega lake system and was located at a sub-tropical latitude (e.g., Xu et al., 2017; Fig. 1A). The occurrence of marine palynomorphs (e.g., *Veryhachium collectum* and *Halosphaeropsis liassica*) and clay minerals (glauconite and chamosite), combined with evidence from geochemistry, suggests that a marine incursion occurred in the Sichuan intracontinental basin during the T-OAE (Xu et al., 2017; Liu et al., 2024). The Lower Jurassic sediments in the northeastern part of the Sichuan Basin constitute the Ziliujing Formation, which is divided into four members. In stratigraphic order, they are the Zhenzhuchong, Dongyuemiao, Ma'anshan, and Da'anzhai members (e.g., Li et al., 2020; Qiu and He, 2022; and reference therein). The Da'anzhai Member mainly consists of dark-grey mudstones/shales with high organic matter content, alternating with abundant fossiliferous limestones (e.g., Cui et al., 2023, and reference therein). A lacustrine transgressive–regressive sedimentary cycle has been identified in the Da'anzhai Member, with the maximum deepening of the lacustrine system occurring around its middle part (e.g., Qiu and He, 2022; Cui et al., 2023). The Ziliujing Formation is overlain by the Lianggaoshan Formation. In some papers (e.g., Qiu and He, 2022), the Lianggaoshan Formation has been referred to as the Xintiangou Fm., as they are contemporaneous deposits but have different facies. In this paper, we followed Cui et al. (2023) in naming the strata overlying the Da'anzhai Member as the Lianggaoshan Formation.

The Dacao 'D' section (DCD) is located ~13 km SW of Dazhou city (31°08'00.0"N, 107°21'49.1"E), eastern Sichuan Basin (Fig. 1B). It consists of the uppermost part of the Ma'anshan Member, the entire Da'anzhai Member, and the lower part of the Lianggaoshan Fm. (Fig. 2). Franceschi et al. (2023) studied two nearby sections, naming them the Dacao 'A' and Dacao 'B' sections, along with the Liangshuijing section (located ~18 km southeast of the Dacao sections). These three sections (Dacao 'A', 'B' and Liangshuijing sections) comprise the upper part of the Zhenzhuchong Member, Dongyuemiao Member, Ma'anshan Member, and the lowermost part of the Da'anzhai Member (Franceschi et al., 2023). Franceschi et al. (2023) reconstructed a composite section from Dacao 'A', 'B' and Liangshuijing sections. Based on carbon isotope stratigraphy and palynostratigraphy, they assigned the stratigraphic interval exposed in the Dacao 'A', 'B' and Liangshuijing sections to upper Sinemurian–lowermost Toarcian. A marked ~8‰ NCIE, coinciding with palynological evidence for a humid climate and lake expansion, was recorded at the inferred Sinemurian–Pliensbachian transition. This was suggested to be the expression in terrestrial environments of a global perturbation in the $\delta^{13}\text{C}$ record known as the Sinemurian–Pliensbachian Boundary Event (e.g., Korte and Hesselbo, 2011; Franceschi et al., 2023).

In this study, we mainly focused on the interval of the Da'anzhai Member in the DCD section (Fig. 2). A preliminary analysis of geochemical detrital proxies shows that the lower part of the Da'anzhai Member in the DCD is characterized by a warm and arid climate, whereas the upper part of the Da'anzhai Member exhibits more humid conditions (Reolid et al., 2024). The petrology of the Da'anzhai Member in the DCD section has been also studied previously, resulting in its division into four subfacies: semi-deep to

deep lake, lake slope, shallow lake, and littoral lake (Zhou, 2021). Cui et al. (2023) have described in detail the sedimentological and petrological features of the Da'anzhai Member from three cores (X1 to X3) located close to the DCD section (Fig. 1B). Therefore, we provide only a brief description of the facies in this paper.

The topmost part of the Ma'anshan Member is mainly composed of greyish-yellow sandstone with interbedded grey finer sediments such as marly sandstones and siltstones, indicative of shallow conditions (lacustrine alluvial plain/delta) (Franceschi et al., 2023). The overlying Da'anzhai Member in the DCD section starts at ~12.2 m (Fig. 2), characterized by thick fossiliferous limestone (~40 cm bed thicknesses) with interbedded sandy mudstones/shales (Fig. 3A, B). The fossiliferous limestones mainly consist of randomly scattered fragments of bivalves and ostracods. This facies association represents a high-energy and shallow water environment, also affected by storm or intense wave action (Cui et al., 2023). Dark grey or black shales/muddy siltstones dominate the middle part of the DCD section (~30 to 65 m) where they alternate with fossiliferous limestones (Fig. 3C, D). These limestone beds are generally thinner than those from the lower part of the section (~11 to 21 m), and often have higher carbonate mud content (Fig. 3E, G), likely indicative of a bioclastic shoal ramp domain (Cui et al., 2023). Thick limestone layers (~60 cm bed thicknesses) are recorded between ~68 and 71 m (Fig. 3F). The uppermost part of the Da'anzhai Member (~99 to 104 m) features thick fossiliferous limestones again (Fig. 3H). These limestone beds are mainly composed of scattered bivalve fragments, suggesting a bioclastic shoal environment characterized by high energy and a relatively low lake level (e.g., Cui et al., 2023). The topmost part of the DCD section (> ~104 m) consists of fine to coarse sandstone layers and is assigned to the lower portion of the Lianggaoshan Formation (Fig. 2). The depositional age of this upper portion remains uncertain. However, a recent astronomical age scale constrained the depositional age of the Lianggaoshan Formation to the early Middle Jurassic (172.74–170.00 Ma) (Lu et al., 2024).

3. Methods

3.1. Palynofacies and palynostratigraphy

A total of 29 samples were selected for palynofacies analysis and palynostratigraphical evaluation. For each sample, about 8–20 g of rock was cleaned and crushed before one tablet containing the exotic marker *Lycopodium* was added (one tablet contains 13,761 *Lycopodium clavatum* spores). The procedures followed the treatment described in Moore et al. (1991), Wood et al. (1996), and Riding (2021), including soaking the samples in sodium pyrophosphate ($\text{Na}_4\text{P}_2\text{O}_7$) for 24 h, followed by acid treatment with 10% HCl and concentrated HF (90 min in a hot water bath at 60 °C, twice repeated) to dissolve carbonate and silicate fractions, respectively. The remaining residue was treated with concentrated HCl to dissolve fluoride minerals. The sample was then washed until neutral, and density separation was performed with heavy liquid ZnCl_2 (specific gravity 2.1 kg/L) to separate the organic material and remaining mineral grains. Residues were washed with deionized water until a neutral pH was reached, cleaned in an ultrasonic bath, and then sieved using a 10 μm nylon mesh and mounted on slides using glycerol. Palynological preparation was performed at the Department of Geology, Croatian Geological Survey. Analysis of the slides was carried out with an Olympus BH2 RFC trinocular transmitted light microscope at magnifications of $\times 100$, $\times 200$ and $\times 400$. Images were taken with an AmScopeTM camera adapter and AmScope v.3.7 software.

At least 300 organic particles from each sample were identified for palynofacies analysis. The palynofacies terminology

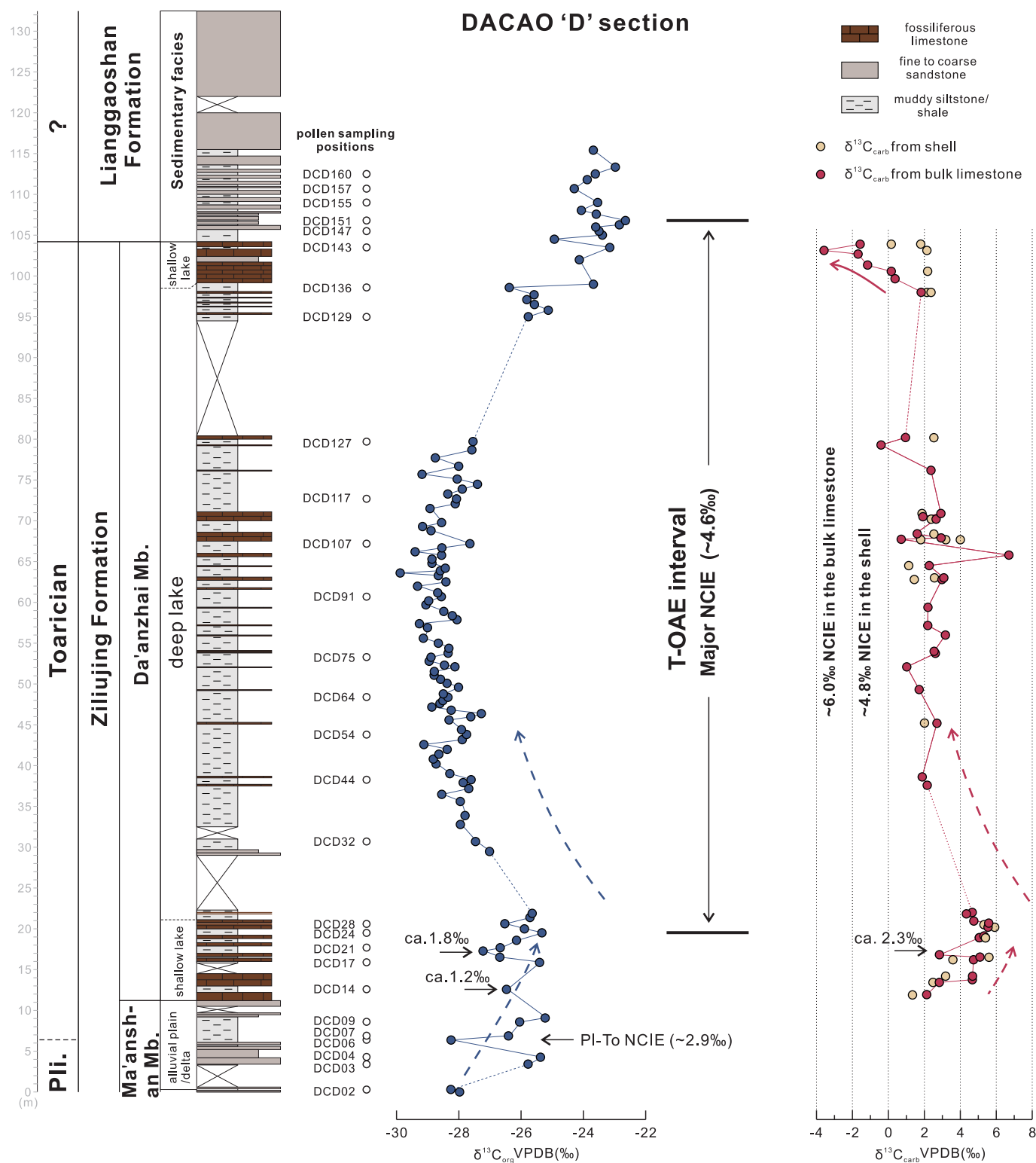


Fig. 2. Lithological composition, sedimentary facies, and carbon isotope data of the Dacao 'D' (DCD) section. T-OAE: Toarcian Oceanic Anoxic Event; NCIE: Negative carbon isotope excursion; Mb.: Member; PI-To: Pliensbachian-Toarcian. The lithological log of the DCD section follows [Zhou \(2021\)](#) and [Reolid et al. \(2024\)](#).

follows [Tyson \(1995\)](#) and [Oboh-Ikuenobe and de Villiers \(2003\)](#). Palynofacies data are presented as percentage proportions based on counting of ≥ 300 particles for each sample ([Fig. 4](#)). Age-diagnostic palynomorph taxa were semi-quantitatively evalu-

ated following the palynostratigraphical schemes described in [Schulz \(1967\)](#), [Filatoff \(1975\)](#), [Van Erve \(1977\)](#), [Lund and Pedersen \(1985\)](#), [Dybkjær \(1991\)](#), and [Koppelhus and Dam \(2003\)](#).

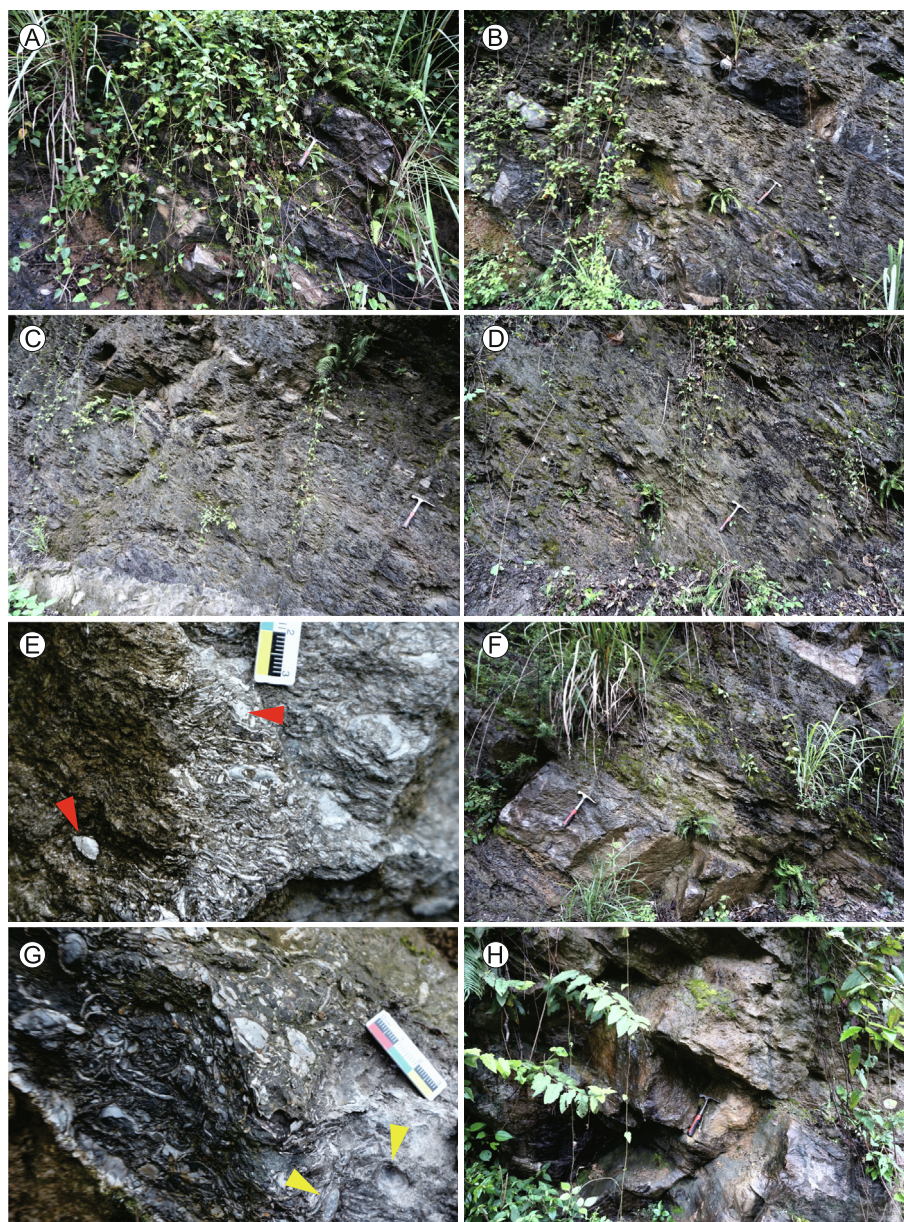


Fig. 3. Field photographs of the DCD section. (A) The thick fossiliferous limestone layers are from the lowest part of the Da'anzhai Member of the DCD section (~14 m in Fig. 2). (B) Fossiliferous limestone layers with interbedded muddy siltstones/shales (~17 m in Fig. 2). (C) The view of back shales with interbedded fossiliferous limestone layer at ~37 m. (D) The middle part of the DCD section is mainly composed of back shales with interbedded thin fossiliferous limestone layers (~54 m in Fig. 2). (E) Close up of fossiliferous limestone layer at ~63 m. The fossiliferous limestone mainly consists of bivalve fragments and calcareous muds. The altered bivalve shells are white in color (red arrows). (F) Two thick fossiliferous limestone layers with interbedded back shales (~69 m in Fig. 2). (G) Close up of fossiliferous limestone layer at ~80 m. The unordered fossils are mainly composed of bivalve fragments which are surrounded by dark grey calcareous muds. Some bivalves are well preserved having smooth and lustrous surfaces (yellow arrows). (H) The topmost part of the Da'anzhai Member mainly consists of thick fossiliferous limestone layers (~101 m in Fig. 2). Hammer is ca. 28 cm long. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

3.2. SEM imaging and cathodoluminescence

A total of six fossiliferous limestones were selected for Scanning Electron Microscope (SEM)-Energy Dispersive X-Ray Spectrometry (EDS) analysis. Bivalves were imaged using a NANO SEM450 instrument in order to evaluate shell preservation. The elemental analysis was performed using EDS (APOLLO X) connected to the SEM. The detection limit is greater than 0.1%. These experiments were performed in the laboratory of the College of Earth and Planetary Sciences at Chengdu University of Technology (CDUT). Cathodoluminescence (CL) was utilized to assess diagenetic alteration in carbonate samples at Beijing Geoanalysis Co. Ltd. Observations of carbonate thin sections were conducted using a cold

cathodoluminescence device. The voltage was set to 12 kV, and the cathode current intensity was maintained at ca. 300 μ A.

3.3. Carbonate carbon and oxygen isotope analyses

A total of 46 bulk limestones and 35 fossil fragments (mainly bivalves) were collected for carbonate carbon isotope ($\delta^{13}\text{C}_{\text{carb}}$) and oxygen isotope ($\delta^{18}\text{O}_{\text{carb}}$) analyses. The shells in the field exhibit two distinct colors: grey and white (Fig. 3E, G). The white parts of a shell result from diagenesis, and they are often loose and can be easily broken by fingernails (Fig. 3E). In contrast, the grey parts of a shell often have smooth and lustrous surfaces and are considered well-preserved in hand specimen (Fig. 3G). Powders were

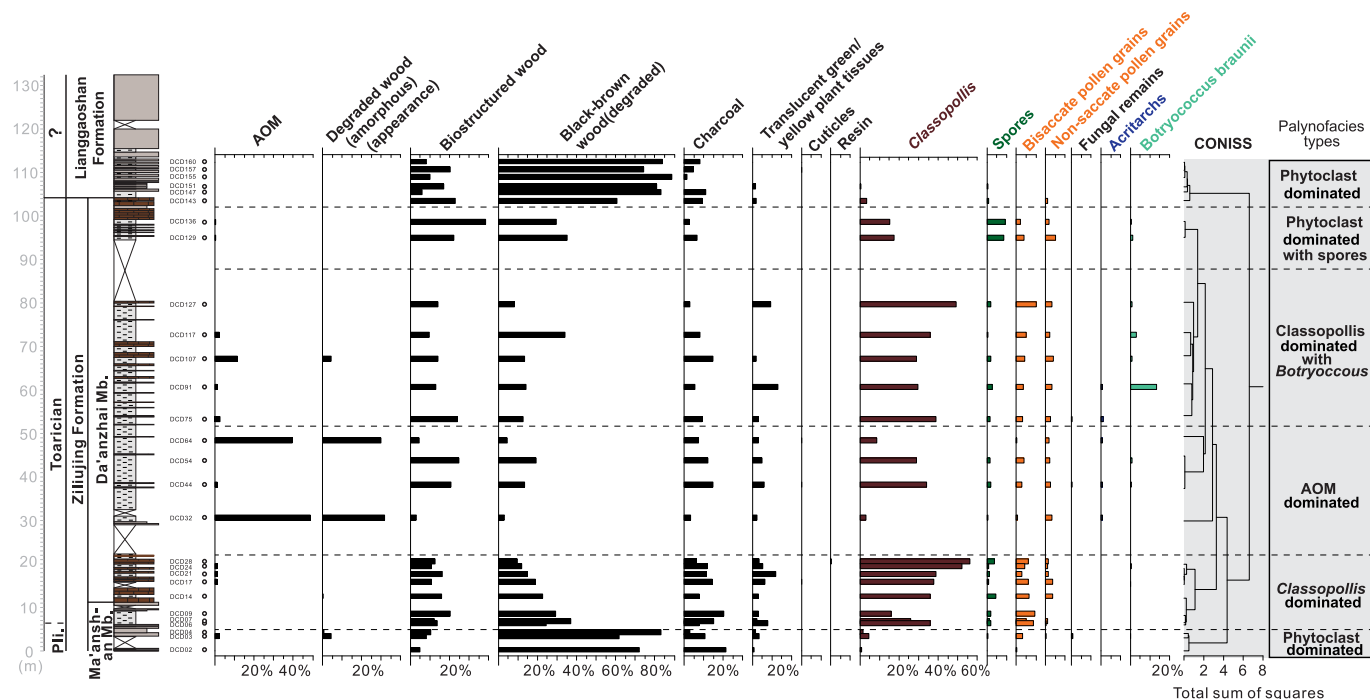


Fig. 4. Palynofacies trends in the DCD section during the T-OAE. AOM: Amorphous organic matter. The group “Non-saccate pollen grains” includes cycadalean, bennettitalean, Cupressaceae-related pollen, and *Sciadopityspollenites* (*Cerebropollenites*). Palynofacies trends are defined as percentage proportion based on counting at least 300 sedimentary organic particles. Palynofacies assemblages are determined by stratigraphically constrained cluster analysis (CONISS, Grimm, 1987).

carefully extracted from the smooth parts of a shell and/or the homogeneous parts of fossiliferous limestones (avoiding calcite veins and bioclasts) using a hand-held drill. These powders were weighed into exetainer vials and reacted with pure phosphoric acid at 70 °C to produce CO₂. The evolved gas was then sent to a Gas-Bench II interfaced with Thermo Finnigan MAT 253 mass spectrometer at the College of Earth and Planetary Sciences at CDUT. The analytical precision (1 σ) for both carbon and oxygen was $\leq 0.12\text{‰}$, as determined through calibration against three international reference standards: GBW04405 ($\delta^{13}\text{C} = 0.58\text{‰}$, $\delta^{18}\text{O} = -8.49\text{‰}$), ANU-M2 ($\delta^{13}\text{C} = 2.81\text{‰}$, $\delta^{18}\text{O} = -7.34\text{‰}$), ANU-PRM2 ($\delta^{13}\text{C} = 0.71\text{‰}$, $\delta^{18}\text{O} = -17.41\text{‰}$), which were repeatedly run along with the samples. The isotope values were calibrated to the VPDB scale. The data are provided in the [Supplementary Data](#). Among the $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ data, 4 of the 46 bulk limestones have been published in the master's thesis of Zhou (2021) and these are highlighted in the [Supplementary Data](#). The $\delta^{18}\text{O}_{\text{carb}}$ values were not corrected using the phosphoric acid fractionation factors proposed by Kim et al. (2007).

3.4. Organic carbon isotope analyses

A total of 118 bulk rock samples were collected for organic carbon isotope analysis. Of these, 101 were analyzed at the Institute of Geochemistry, Chinese Academy of Sciences (IGCAS) in Guiyang, China, and 61 were analyzed at the Key Laboratory of Submarine Geoscience, Second Institute of Oceanography (Hangzhou, China). To ensure data quality, 44 of the 118 samples were analyzed at both laboratories for cross-validation. The 29 samples for the palynofacies analysis were selected from among these 118 bulk rock samples. Before isotope analysis, fresh bulk rocks were selected and cleaned with deionized water and then oven-dried at 55 °C for several hours. After drying, samples were crushed to a size of 200 μm using a tungsten grinder. Approximately 2 g of powder was weighed into a polypropylene Falcon tube, to which 10% HCl was added to remove carbonate. The residues were then cen-

trifuged and repeatedly rinsed with deionized water until the water reached a neutral pH. The residues were further dried in an oven at 55 °C for ~ 24 h. The wet chemical preparations were conducted in the College of Material and Chemistry & Chemical Engineering, CDUT. In Guiyang, the dried decalcified powders were placed into tin capsules and analyzed using a Thermo Finnigan MAT 253 isotope ratio mass spectrometer linked to a Thermo Finnigan Flash EA 2000. Results are reported in delta notation relative to the Vienna Pee Dee Belemnite (VPDB). The international standard IAEA-C3 cellulose was measured repeatedly during the analysis. The analytical precision obtained was better than $\pm 0.1\text{‰}$ (1 σ). In Hangzhou, the decalcified powders were analyzed using a Thermo Delta plus Advantage isotope ratio mass spectrometer connected to a Flash EA 1112 analyzer. Two reference standards, USGS-24 (-16.05‰) and GBW4408 (-36.9‰) were analyzed with the samples. The analytical precision achieved was better than $\pm 0.2\text{‰}$ (1 σ). The data are presented in the [Supplementary Data](#). Among the organic carbon isotope data, 12 of the 118 bulk rock samples have been previously published in the master's thesis of Zhou (2021), and these are highlighted in the [Supplementary Data](#).

4. Results

4.1. Palynofacies and palynostratigraphy

The sedimentary organic matter (SOM) in the DCD section is primarily of terrestrial origin. Opaque black-brown phytoclasts (up to 87% of the palynofacies counts) are the most abundant palynofacies group, which includes all oxidized brownish black to black woody tissues, or carbonized particles of plant-derived kerogen. Non-opaque biostructured phytoclasts including wood and tracheid particles comprise up to 41% of the sedimentary organic matter. Amorphous particles including resin and amorphous organic matter (AOM, structureless dispersed organic particles of terrestrial and/or aquatic origin) are rare to less common (abundance

usually less than 10 counts). *Classopollis* is the major component among palynomorphs (up to 61% of total kerogen particles) followed by spores and bisaccate pollen grains (each up to ~10% of the SOM particles). Marine palynomorphs, acritarchs and freshwater algae (*Botryococcus braunii*) are recorded rarely (<5 counts) with the exception of the interval between 60.7 and 72.7 m, where *Botryococcus* abundances reach up to ~16% of the total kerogen particles.

Five major palynofacies types are identified in the DCD section (Fig. 4) using stratigraphically constrained cluster analysis (CONISS) built in Tilia/Tilia-Graph (Grimm, 1987). Phytoclasts, black brown and biostructured wood particles are dominant in the lowermost 4 m of the section (0.3–4.3 m) as well as in its uppermost part (103.5–112.5 m) (Fig. 4). A sharp rise in *Classopollis* abundance is recorded at 6.4 m (~38%) followed by overall high values with some oscillation in the interval of 6.4–20.6 m (Fig. 4). The palynofacies between 30.7 m and 48.4 m are characterized by high abundance of amorphous organic matter (>40%) and a decrease in the phytoclast group and palynomorphs (Fig. 4). In the same interval a pronounced drop occurs in the abundance of bisaccate and non-saccate pollen from ca. 10% to less than 5% of the SOM particles. The abundance of phytoclasts and palynomorphs rises again at 53.5–79.7 m (Fig. 4) with distinct horizons of increased *Botryococcus braunii* abundances particularly in samples DCD 91 (60.7 m) and DCD 117 (72.7 m). Higher in the section, bisaccate pollen grains (and to a lesser extent non-saccate pollen grains) increase in abundance. The interval between 95.0 m and 98.6 m is characterized by an overall increase in phytoclast and palynomorph abundances, as well as an increase in the diversity of palynomorphs. Spore abundance also increases (~8%–10% of SOM), whereas *Classopollis* abundance falls relative to the lower part of the section, from >40% down to ~20%. The topmost part

of the succession has an impoverished palynofacies with mainly oxidized black-brown and biostructured plant debris (Fig. 4).

The semi-quantitative distribution of selected palynostratigraphically important taxa is illustrated in Fig. 5. *Classopollis* species are predominant through the entire succession. Age-diagnostic taxa (Schulz, 1967; Filatoff, 1975; Van Erve, 1977; Lund and Pedersen, 1985; Dybkjær, 1991; Koppelhus and Dam, 2003) characteristic for the Toarcian, *Ischyosporites variegatus* and *Contignisporites problematicus* appear first at 6.4 m. *Leptolepidites macroverrucosus* is recorded first at 6.9 m. *Sciadopityspollenites (Cerebropollenites) macroverrucosus* is present from 15.9 m to 95 m in the DCD section with more frequent occurrences at 15.9–19.5 m. The semiquantitative distribution of biostratigraphically significant taxa is provided in the Supplementary Data.

4.2. SEM and CL of the lacustrine carbonates

The fossiliferous limestones (e.g., shell limestones and muddy shell limestones) are mainly composed of bivalve shells, microsparry calcite cement and carbonate mud (Fig. 6A, C). Most of the bivalve shells are fragmented. Some show evidence of recrystallization (Fig. 6A), displaying dull or no CL (Fig. 6B). Some shells retain the original shell structures, as indicated by the presence of foliated layers and growth lines (Fig. 6C), and exhibit a homogeneous dark-blue CL (Fig. 6D). Interparticle spaces are filled with microsparry calcite cement (or fine dolomite crystals) that exhibit dull or no CL (Fig. 6C, D). Carbonate mud filling these spaces shows heterogenous dull to medium CL (reddish-orange colors in Fig. 6C, D).

Under the SEM, the well-preserved shells exhibit a smooth, foliated structure (Fig. 6E, F), whilst some shells were completely dissolved and replaced by blocky calcite crystals (Fig. 6G). The altered shells consistently display an etched texture of calcite fibres (Fig. 6H).

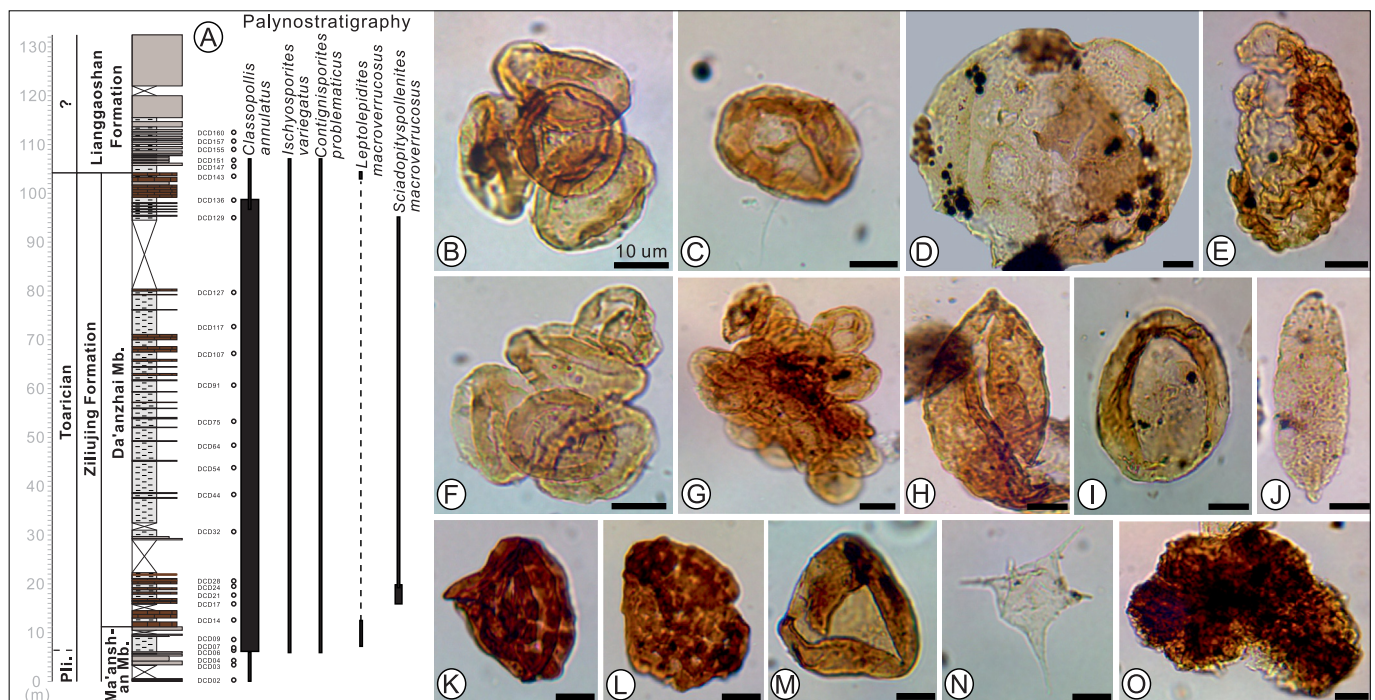
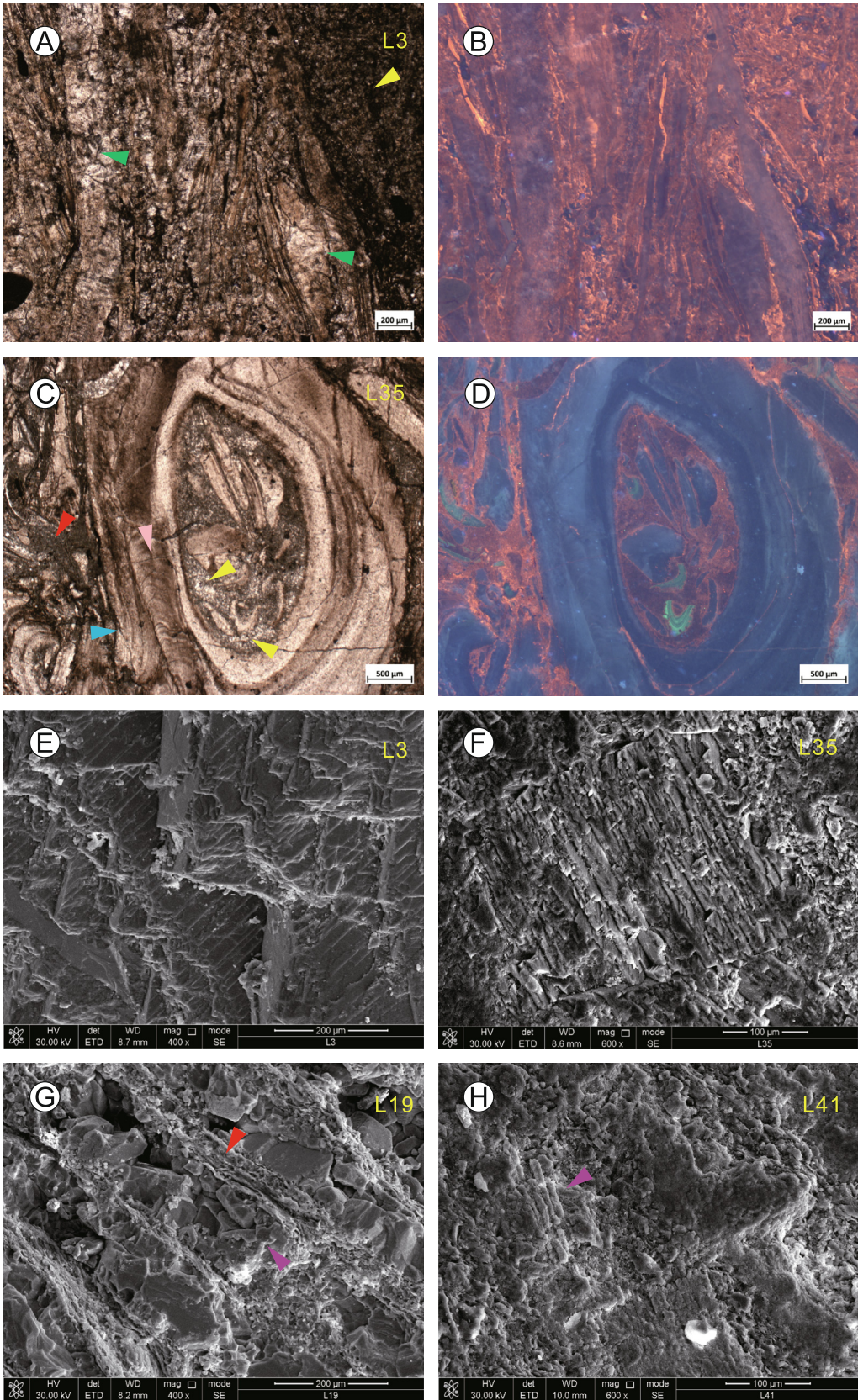


Fig. 5. Palynostratigraphical constraints (A) and representatives of the main palynomorph groups (B–O) identified during the palynofacies analysis from the DCD section. The scale bar represents 10 μ m. (B) *Classopollis annulatus* tetrad, DCD 21. (C) *Classopollis annulatus* single grain, DCD 03. (D) *Alisporites pergrandis*, DCD 21. (E) *Sciadopityspollenites (Cerebropollenites) macroverrucosus*, DCD 21. (F) *Classopollis annulatus* tetrad, DCD 75. (G) *Spheripollenites* sp. agglomerate, DCD 03. (H) *Chasmatosporites hians*, DCD 107. (I) *Chasmatosporites apertus*, DCD 54. (J) *Cycadopites typicus*, DCD 24. (K) *Contignisporites problematicus*, DCD 07. (L) *Ischyosporites variegatus*, DCD 07. (M) *Cyathidites mesozoicus*, DCD 28. (N) *Verychahium* sp., DCD 54. (O) *Botryococcus braunii*, DCD 24.



4.3. Carbonate carbon and oxygen isotopes of lacustrine carbonates

Carbonate carbon and oxygen isotopes (ten from shell and two from limestone) with a high standard deviation (>0.1) in the carbon isotope value due to low sample weight are discarded. Carbon and oxygen isotopes from bulk limestones exhibit $\delta^{13}\text{C}_{\text{carb}}$ values between -3.58‰ and 6.70‰ (avg. = 2.49‰ ; $n = 44$), and $\delta^{18}\text{O}$ values between -16.20‰ and -4.82‰ (avg. = -12.37‰). Shells show $\delta^{13}\text{C}_{\text{carb}}$ values from 0.16‰ to 5.91‰ (avg. = 2.76‰ ; $n = 25$), and $\delta^{18}\text{O}$ values ranging from -16.19‰ to -5.61‰ (avg. = -10.66‰).

$\delta^{13}\text{C}_{\text{carb}}$ values of shells show a positive shift in the interval between ~ 11.9 and ~ 20.2 m within the lower part of Da'anzhai Member. In contrast, within this interval, $\delta^{13}\text{C}_{\text{carb}}$ values of bulk limestones display a negative shift (ca. -2.3‰). Above this (~ 20.2 to ~ 98 m), a more protracted negative $\delta^{13}\text{C}_{\text{carb}}$ shifts occur in both bulk limestones and shells, with magnitudes of ca. -6.0‰ and ca. -4.8‰ (Fig. 2).

4.4. Organic carbon isotopes of bulk rocks

In general, $\delta^{13}\text{C}_{\text{org}}$ values from the Guiyang laboratory range from -29.88‰ to -22.87‰ , with an average of -27.32‰ ($n = 101$), and $\delta^{13}\text{C}_{\text{org}}$ values from the Hangzhou laboratory range from -29.10‰ to -22.68‰ , with an average of -26.85‰ ($n = 61$). Student's t -test results indicate no significant difference between the Guiyang and Hangzhou datasets (see Supplementary Data). Therefore, in the subsequent analysis, $\delta^{13}\text{C}_{\text{org}}$ data from the two laboratories have been merged by averaging the two values for each of the 44 samples (see Supplementary Data).

A positive trend in $\delta^{13}\text{C}_{\text{org}}$ values starts at the base of the DCD section that persists up to ~ 19.5 m. This trend includes a sharp NCIE at the topmost part of the Ma'anshan Member (ca. -2.9‰), followed by two minor fluctuations (ca. -1.2‰ and -1.8‰) observed in the lowermost part of the Da'anzhai Member. Subsequently, a broad negative excursion, referred to as the major NCIE, begins at ~ 19.5 m and ends at ~ 106.8 m, with a magnitude of ca. -4.6‰ . This major NCIE spans almost the entire Da'anzhai Member (Fig. 2).

5. Discussion

5.1. Age constraints from palynology for the Pliensbachian–Toarcian boundary and T-OAE interval in the Sichuan Basin

The predominance of *Classopollis* pollen characterizes the T-OAE interval worldwide (e.g., Vakhrameyev, 1991; Batten and Koppelhus, 1996; Wang et al., 2005; Zakharov et al., 2006; Ashraf et al., 2010; Correia et al., 2018; Slater et al., 2019; Jin et al., 2020; Galasso et al., 2022; Baranyi et al., 2024) and sediments of this age can be assigned to the Australasian *Classopollis torosa* Oppel Zone (Helby et al., 1987). The first significant rise in *Classopollis* abundance is usually recorded at the Pliensbachian–Toar-

cian boundary likely coinciding with the onset of a global climate warming phase (e.g., Slater et al., 2019). This is followed by a decline in the pre-TOAE NCIE interval due to a cooling episode in the earliest Toarcian, possibly accompanied by the development of ice caps (Suan et al., 2010; Dera et al., 2011; Ruebsam et al., 2019; Baranyi et al., 2024). A similar trend can be observed in the DCD section, with very low *Classopollis* proportions in the lowermost 5 m of the section, followed by an initial rise at 6.4 m (coincident with a sharp NCIE) that we infer to be the Pliensbachian–Toarcian boundary (Figs. 4 and 5). This peak is followed by lower *Classopollis* ratios at 6.9–8.6 m. Between 12.6 and 95 m the palynological assemblages are dominated by *Classopollis annulatus* and to a lesser extent by *C. quiyangensis*, in accordance with an early Toarcian age (Wang et al., 2005). This interval is also marked by major NCIEs in both bulk rock and lacustrine carbonates (Fig. 2).

Despite the fact that the majority of the palynomorph taxa in the DCD section are long-ranging Mesozoic taxa that existed from the Triassic to the Cretaceous (e.g., Barrón et al., 2010), some species provide further support for the early Toarcian age interpretation in our samples. In the DCD section, the local first occurrence of both *I. variegatus* and *C. problematicus* is at 6.4 m, and coincident with the sharp NCIE at the topmost part of the Ma'anshan Member that marks the Pliensbachian–Toarcian boundary (Figs. 2 and 4). These species appeared first in the latest Pliensbachian but became more common in the early Toarcian in Europe and Gondwana (Schulz, 1967; Herengreen and De Boer, 1974; Filatoff, 1975; Van Erve, 1977; Koppelhus and Dam, 2003). Similarly, *Leptolepidites macroverrucosus*, which was recorded first at 6.9 m in the DCD above the proposed boundary, first appeared in the Toarcian in Europe (Barrón et al., 2010).

Agglomerates of *Spheripollenites* pollen were found in samples from 38.3 to 53.3 m. European (and Boreal) early Toarcian palynological zonations have often denoted the predominance of *Spheripollenites subgranulatus* or *S. psilatus* (e.g., Morbey, 1978; Lund and Pedersen, 1985; Dybkjær, 1991; Koppelhus and Batten, 1996) often forming the majority of palynological assemblages ($>75\%$). Some of these palynomorphs were indeed gymnosperm pollen probably belonging to the Cupressaceae (Guy-Ohlson, 1986), but it has to be noted that the majority of them were marine algae, prasinophytes labelled informally as “sphaeromorphs” (or *Halosphaeropsis liassica*) characteristic of the organic rich sediments that span the T-OAE NCIE (e.g., Prauss and Riegel, 1989; Prauss et al., 1991; Bucefalo Palliani and Riding, 1999; Bucefalo Palliani et al., 2002).

In a series of boreholes in Øresund (Denmark) a *Spheripollenites-Leptolepidites* Zone was defined and dated to the Toarcian, which was characterized by the first appearance of *I. variegatus* and *Manumia delcourti* (Dybkjær, 1991; Koppelhus and Batten, 1996). Assemblage B of Lund and Pedersen (1985) from East Greenland is also characterized by the appearance of *I. variegatus* and *Contignisporites* spp. in the Toarcian. The same palynological assemblage was also characterized by the common occurrence of *Sciadop-*

Fig. 6. Thin-section, CL, and SEM microphotographs of lacustrine carbonates. (A) Limestone sample L3 (~ 13.4 m in Fig. 2). Bioclasts are mainly composed of bivalve fragments. Some bivalves are partly altered and recrystallized by blocky calcites (green arrows). Interstitial spaces are mainly filled with microsparry calcites and carbonate muds (yellow arrows). (B) CL image of A, altered shells exhibit dull or no CL. The microsparry calcites and muds display dull to medium CL. (C) Limestone sample L35 (~ 68.4 m in Fig. 2). Fragments of bivalves are dominated. Interstitial spaces are filled with microsparry calcites (or fine dolomite crystals) (yellow arrows) and carbonate muds (red arrow). The well-preserved bivalves are originally made of aragonites. The foliated layer (blue arrow) and growth line (pink arrow) of bivalve are visible. (D) CL image of C, the well-preserved bivalves display dark-blue CL. The microsparry calcites/dolomites show dull or no CL, while the carbonate muds exhibit heterogeneous dull to medium CL. (E) SEM image of bivalve shell from sample L3. The unaltered bivalve shell displays a smooth, foliated structure. (F) SEM image of bivalve shell from sample L35. Another visual angle shows the foliated structure of the well-preserved bivalve shell. (G) SEM image of bivalve shell from limestone sample L19 (~ 38.6 m in Fig. 2). The bivalve shells in this visual field have been completely dissolved, leaving only the shells' mold, which have been occupied by late calcite crystals (purple arrow). Interparticle spaces are filled with carbonate muds (red arrow). (H) SEM image of bivalve shell from limestone sample L41 (~ 80.2 m in Fig. 2). In this visual field, bivalve shells are partly altered, characterized by the etched textures of foliated structures (purple arrow). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

tyspollenites (*Cerebropollenites*) *macroverrucosus* (Kelly et al., 2015). This species is also common in the DCD samples, particularly at 15.9–19.5 m.

The minimum (youngest) age of the section cannot be younger than Toarcian due to the lack of the characteristic Middle Jurassic *Callialasporites* 'complex' (Van Erve, 1977; Srivastava, 1987; Rauscher and Schmitt, 1990; Quattrocchio et al., 1996; Quattrocchio et al., 2011). *Chasmatosporites* is very common in the entire DCD section (see non-saccate pollen grains in Fig. 4 and the Supplementary Data). Species of this genus are characteristic for the Early Jurassic in Europe and rarer in sediments younger than Toarcian (Batten and Koppelhus, 1996; Xu et al., 2017).

In summary, the systematic patterns of *Classopollis* relative abundances indicative of regional to global trends, integrated with the presence of key taxa like *I. variegatus*, *C. problematicus*, *Leptolepidites*, *S. macroverrucosus* and the lack of *Callialasporites*, can constrain the age of the DCD succession. Together, these data suggest that the stratigraphic interval examined in this study spans the Pliensbachian–Toarcian transition, and that the age of the youngest strata in the section cannot be younger than Toarcian.

5.2. Petrology of lacustrine carbonates: Diagenetic evaluation and interpretation

Cathodoluminescence (CL) is an important technique for studying carbonate materials (e.g., Marshall, 1988; Barbin, 2000; Boggs and Krinsley, 2010). The CL of carbonate minerals is mainly influenced by the relative concentrations of manganese (Mn), rare-earth elements (REEs), and iron (Fe). The of Mn^{2+} is the main activator of extrinsic CL, whereas Fe^{2+} serves as an important quencher (e.g., Boggs and Krinsley, 2010; Barbin, 2013, and reference therein). Bivalve shells are either entirely composed of calcite, entirely of aragonite, or a combination of both, with layers of calcite alternating with aragonite (Flügel and Munnecke, 2010). The modern biogenic aragonite and calcite often show different CL colors, typically bright orange for calcite and dark greenish-blue or green for aragonite (e.g., Götte and Richter, 2009; Barbin, 2013). The different intrinsic CL colors in biogenic carbonates (aragonite and calcite) are related to their different crystal lattices (e.g., Götte and Richter, 2009). It is worth noting that some well-preserved marine brachiopod shells also exhibit no CL (e.g., Jin et al., 2018). This absence of CL may be attributed to low Mn contents, as demonstrated in a case study of Silurian brachiopods (Wenzel, 2000).

In the Da'anzhai Member, some bivalve shells that still preserve visible growth lines (Fig. 6C, D), combined with evidence (e.g., smooth surface) from SEM of the shells (Fig. 6E, F), suggest that they were either unaltered or only minimally affected by diagenesis. The well-preserved bivalve shells exhibit homogeneous dark-blue CL (Fig. 6D), which differs from that of altered shells or microsparry calcite/dolomite (dull or no CL) and carbonate mud (dull-medium CL) (Fig. 6B, D), suggesting that the carbonate composition of well-preserved bivalve shells must consist of only one mineral (either calcite or aragonite) and be different from that of the others. XRD analysis of unaltered bivalve shells shows that the carbonate minerals are mainly composed of calcite and aragonite (Supplementary File Fig. S1). The presence of calcite is related to contamination that occurred during the peeling of thin shells from the bulk rocks. The presence of clay minerals further supports this point. Therefore, we conclude that the bivalve shells from the Da'anzhai Member selected in this study are originally composed of aragonite. However, due to the limited petrographic analyses, we cannot fully rule out the possibility that some shells may be originally composed of calcite.

Some bivalve shells have been altered by blocky calcite and no longer retain any original structure (Fig. 6A and G), showing dull or

no CL (Fig. 6B). It is impossible to determine when this alteration occurred, as diagenetic fluids are ubiquitous throughout many diagenetic processes (e.g., eogenesis, mesogenesis, and telogenesis). In these processes, aragonite and high-Mg calcite are easily transformed into low-Mg calcite, which is traditionally regarded as the more stable carbonate mineral (e.g., Flügel and Munnecke, 2010, and references therein). The dull or absent CL in blocky calcite may be related to the relative concentrations of Mn^{2+} and Fe^{2+} , which are linked to changes in redox conditions during different depositional and diagenetic processes, and also other factors (e.g., Machel, 2000; Boggs and Krinsley, 2010).

The interstitial carbonate matrix (e.g., carbonate mud) in our limestones mainly exhibits heterogeneous dull to medium CL (Fig. 6D), possibly suggesting that the concentrations of Mn^{2+} , Fe^{2+} , or rare earth elements are also heterogeneous in the mud. SEM-EDS analysis of the carbonate mud shows the presence of Fe (4.81 wt.%, Supplementary File Fig. S2) which could lead to low-intensity CL. Studies have also shown that 100 ppm (0.01 wt.%) of Fe concentrations is enough to quench CL when Mn is below 200 ppm (e.g., Budd et al., 2000).

The above discussion indicates that unaltered bivalve shells retain the original carbonate mineral that formed in the lake water column. Therefore, they may preserve the original geochemical composition of the lake water. However, the carbonate matrix from the fossiliferous limestones (e.g., shell limestones, muddy shell limestones) of the Da'anzhai Member typically consists of carbonate mud, microspar, and/or fine-grained dolomite, and fine-grained skeletal fragments, which may have undergone different diagenetic processes. However, bulk carbonates of the Da'anzhai Member exhibit low porosity (e.g., Su et al., 2020; Xu et al., 2020, and references therein) and lack features such as blocky calcites or vugs, suggesting that they experienced limited diagenesis.

The fossiliferous limestones of the Da'anzhai Member are characterized by horizontal bedding with different thickness in the field (see in Fig. 3). They are also widely distributed in a large area in the Sichuan Basin (e.g., Qiu and He, 2022, and references therein). This suggests that these carbonate-rich beds cannot be formed within sediment pores but were instead deposited in the water column. Most of these carbonates originated from biogenic sources, including bivalves, ostracods, and gastropods (e.g., Cui et al., 2023), and the original mineral compositions of bivalve shells are mostly aragonitic (e.g., Su et al., 2020). The carbonate-rich beds of the Da'anzhai Member mainly formed under storm conditions in a shallow lake setting (e.g., Cui et al., 2023; Liu et al., 2022a), and some of them were transported to deeper areas by landslides or other processes (e.g., Habulashenmu et al., 2024).

The evidence above suggests that the fossiliferous limestones of the Da'anzhai Member may still preserve part or all of the original geochemistry of the water from the Sichuan paleolake.

5.3. Interpretation of carbon and oxygen stable isotopes from Toarcian lacustrine carbonate in the Sichuan Basin

Studies have suggested that a significant positive correlation between $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ values in lake carbonates hints those carbonates precipitated under conditions of relatively long residence times and stable hydrologic conditions, indicative of a closed lake system (e.g., Talbot, 1990; Anadón and Utrilla, 1993; Benavente et al., 2021). Conversely, a low or non-existent correlation between $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ implies that the carbonates formed in lakes with short residence times, characteristic of more open lake systems (Talbot, 1990; Benavente et al., 2019). Li et al. (2020) observed pronounced positive covariation between $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ values from lacustrine carbonates (e.g., marlstone and dolomitic mudstone; $r^2 = 0.96$; $n = 10$) and pedogenic carbonates (e.g., calcrete; $r^2 = 0.57$; $n = 8$) in the Da'anzhai Member (Fig. 7),

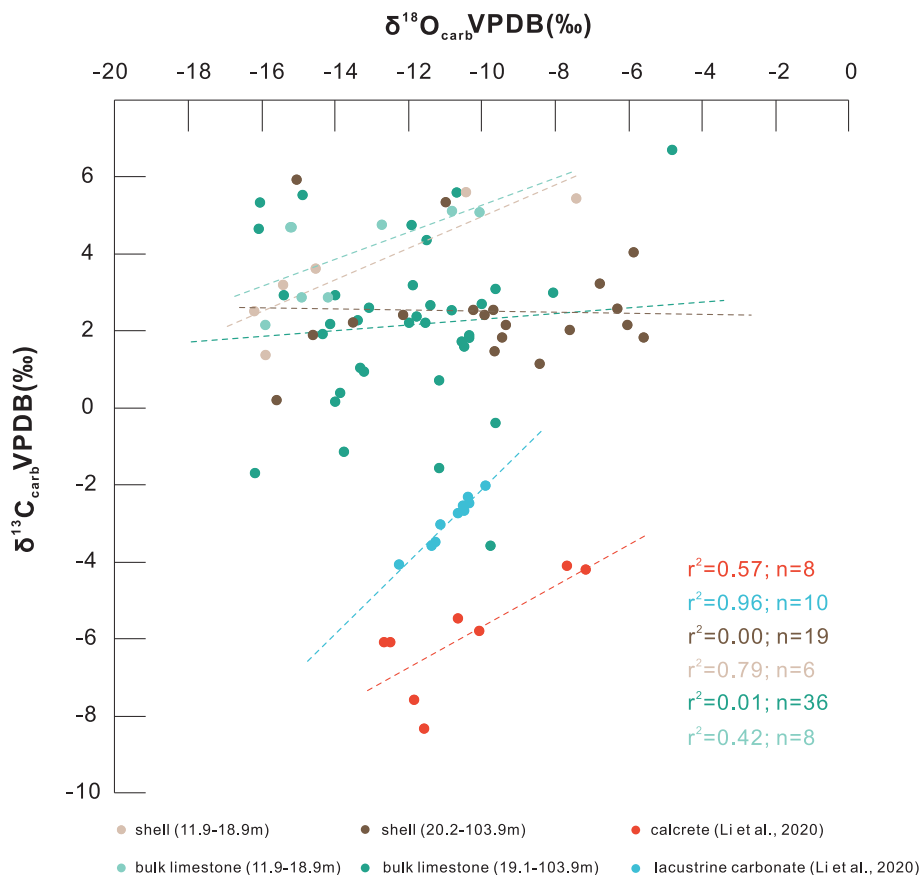


Fig. 7. Correlations between the carbonate carbon and oxygen isotope values of lacustrine carbonates of the Da'anzhai Member. The isotope data for calcretes and lacustrine carbonates are exclusively from the Da'anzhai Member (Li et al., 2020).

suggesting that the Sichuan paleolake was a semi-closed basin during the Early Jurassic. Furthermore, the generally low $\delta^{18}\text{O}_{\text{carb}}$ values of the lacustrine carbonates (ca. -5‰ to -12‰), coupled with negative $\delta^{13}\text{C}_{\text{carb}}$ values between -1‰ and -3.5‰ , imply that the lacustrine system was likely brackish (Li et al., 2020). This evidence is consistent with a regional arid or semi-arid climate where evaporation exceeds precipitation (Li et al., 2020). In contrast, our new $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ data from both the bulk limestones and shells show no correlation (Supplementary File Fig. S3), suggesting the Sichuan paleolake may have been an open system during the Toarcian. However, focusing solely on samples (bulk limestone and shell; 11.9–18.9 m) from the pre-TOAE interval in the DCD section, moderate to strong correlations between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ are observed ($r^2 = 0.42$ in the bulk limestone; $r^2 = 0.79$ in the shell; Fig. 7). This allows us to propose an alternative interpretation: prior to the T-OAE, the Sichuan paleolake was likely semi-enclosed, whereas during the T-OAE the lake was relatively open. This is in line with the observation during the T-OAE implying that a transgression occurred in the Sichuan Basin, coupled with marine incursion, as evidenced by the presence of marine palynomorphs (acritarchs, Figs. 4 and 5) in the palynofacies data and green clay minerals (e.g., glauconite and chamosite) (Xu et al., 2017; Liu et al., 2024), indicating a relatively open and deep lake system.

5.4. The Early Jurassic NCIEs in lacustrine sediments from the Sichuan Basin and their global correlations

The $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ values from the DCD section both exhibit positive trends prior to the T-OAE interval (Fig. 2). This positive shift has also been observed in the carbon isotope composition of

fossil wood and bulk carbonate from marine T-OAE sections in Europe (e.g., Hesselbo et al., 2007; Hermoso et al., 2012), as well as in bulk organic matter from lake sediments in the Ordos Basin of China (e.g., Jin et al., 2020). As alluded to earlier, based on the palynological assemblages the NCIE at the top of the Ma'anshan Member can be linked to the Pliensbachian–Toarcian Boundary Event which is also characterized by a negative carbon isotope excursion (PI-To NCIE) observed in different substrates and paleogeographic settings (e.g., Littler et al., 2010; Fantasia et al., 2019; Kemp et al., 2022, also see Fig. 2). The negative carbon isotope shift, which we correlate with the PI-To NCIE identified in the DCD section, may correspond to the LSJ/n-CIE 2 (equals to the n-CIE4 in the composite carbon isotope curve) recognized by Franceschi et al. (2023) in the uppermost part of the Ma'anshan Member of the Liangshuijing section (~ 18 km southeast of the Dacao section), although there are no robust biostratigraphic constraints available from the Liangshuijing section (Franceschi et al., 2023). Additionally, the PI-To NCIE from the DCD section can be tentatively correlated with the negative perturbations in $\delta^{13}\text{C}_{\text{org}}$ data that are recorded in two other lacustrine cores (Core A and LQ104X) from the Sichuan Basin, despite the lack of solid biostratigraphy (Xu et al., 2017; Liu et al., 2020a; also see in Fig. 8).

Above the inferred PI-To, significant NCIEs of ca. -6‰ and -4.8‰ occurred in $\delta^{13}\text{C}_{\text{carb}}$ in both the bulk limestone and shell, respectively, coincident with a -4.6‰ NCIE in bulk organic matter (Fig. 2). The coupled negative excursions in both carbonate and organic $\delta^{13}\text{C}$ suggest a major change in the carbon cycle occurred, and that this shift is not attributable to diagenetic alterations. The NCIEs span ~ 87 m. The palynostratigraphical data enable the precise correlation of these thick NCIEs in $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{org}}$ to the

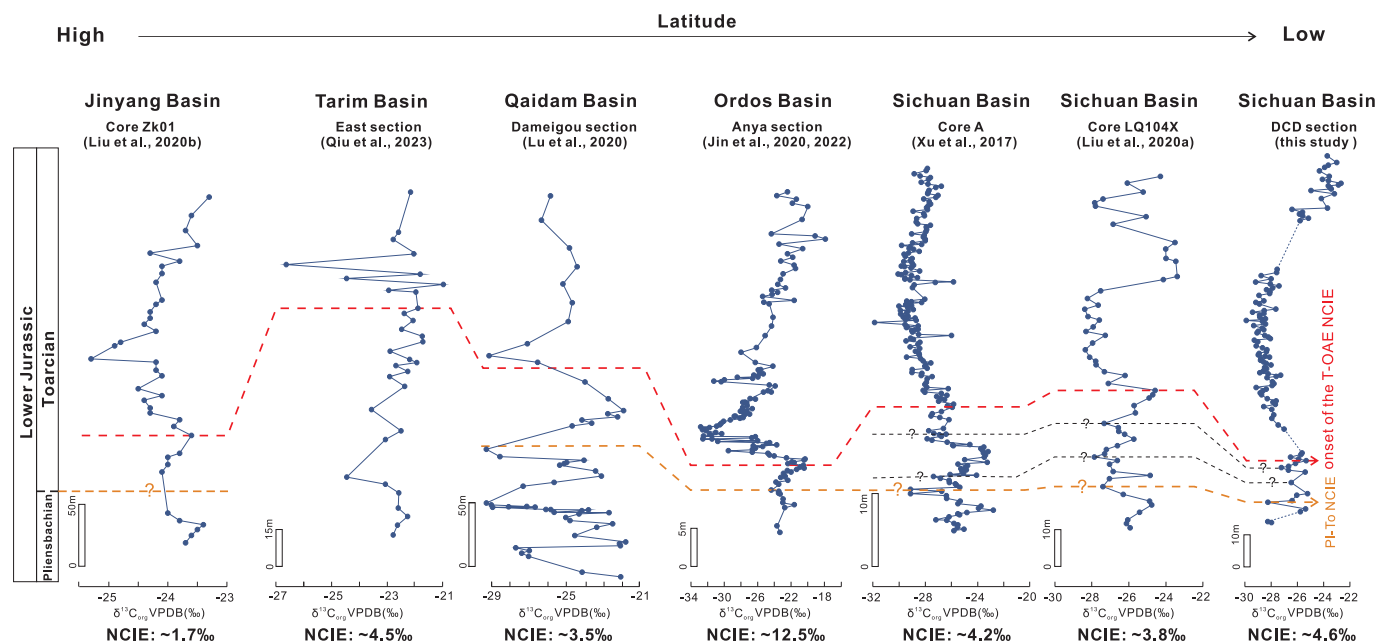


Fig. 8. Comparison of the typical organic carbon-isotope records spanning Pliensbachian–Toarcian stages from Chinese lacustrine basins. The magnitudes of the T-OAE NCIE vary across basins at different latitudes, but within the Sichuan Basin, they are nearly consistent. The definitions and amplitudes of the T-OAE NCIEs from Jinyang, Tarim, Qaidam, Ordos, and Sichuan basins are from Xu et al. (2017), Jin et al. (2020, 2022), Liu et al. (2020a, 2020b), Lu et al. (2020), Qiu et al. (2023) and this study, respectively. The sequence of latitudes for Chinese terrestrial basins follows Deng et al. (2017).

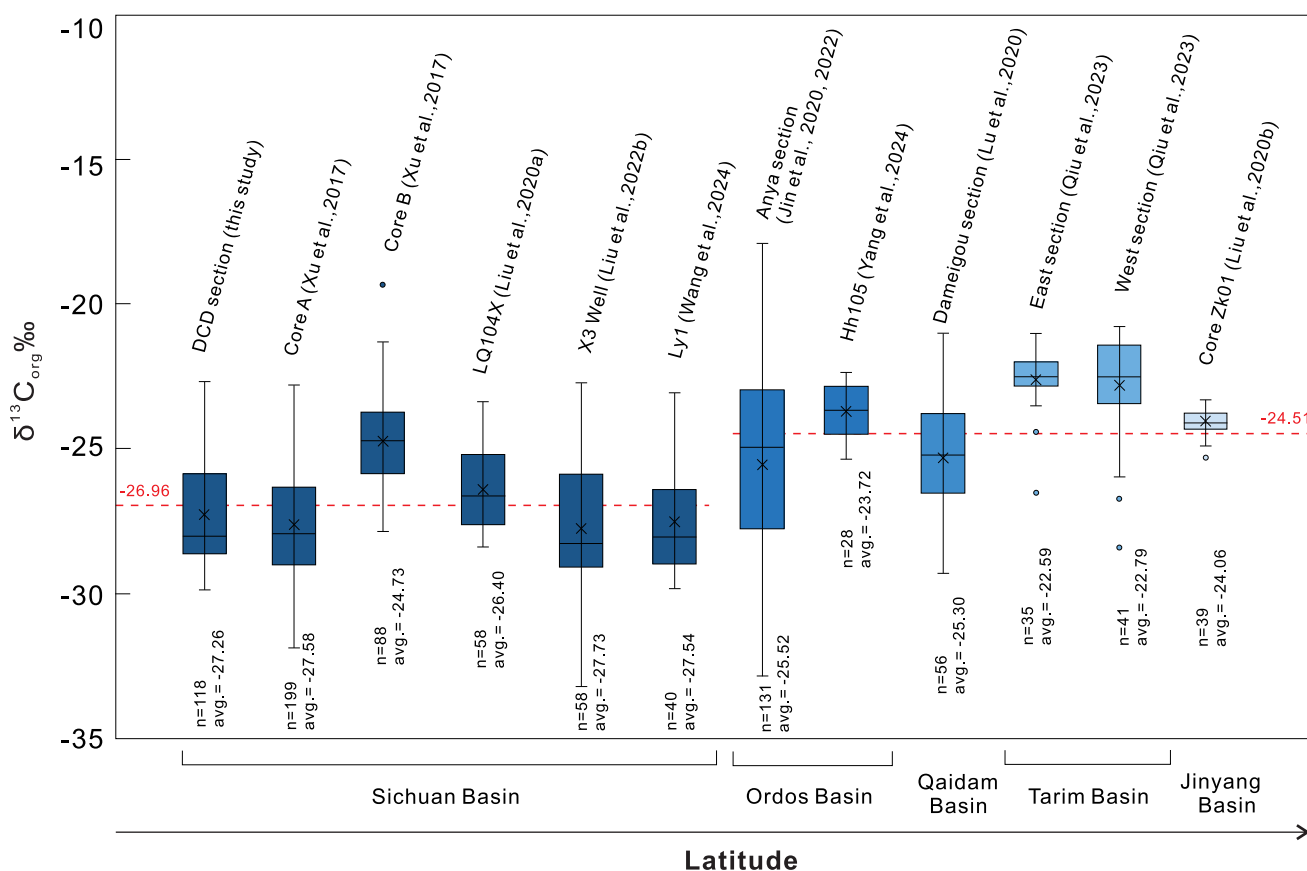


Fig. 9. Box plot of organic carbon-isotope records spanning Pliensbachian–Toarcian stages from Chinese terrestrial basins. The carbon isotope values of the Sichuan Basin (lower latitude; Xu et al., 2017; Liu et al., 2020a, 2022b; Wang et al., 2024; this study) have lower $\delta^{13}\text{C}$ values than those from basins at higher latitudes. The red dashed lines represent the average values of carbon isotopes from low latitude Sichuan basin and high latitude terrestrial basins including Ordos (Jin et al., 2020, 2022; Yang et al., 2024), Qaidam (Lu et al., 2020), Tarim (Qiu et al., 2023), Jinyang (Liu et al., 2020b). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

globally recognized T-OAE NCIE (Figs. 2 and 8). The carbon isotope composition and the magnitude of the T-OAE NCIE data from the DCD section are similar to those from core sediments (Core A) in Sichuan Basin (Xu et al., 2017; Figs. 8 and 9).

Recent high-precision geochronological data obtained from the Panthalassa Ocean indicate that the duration of the T-OAE NCIE lasted at least 169 kyr but less than 407 kyr (Kemp et al., 2024). If this age model for the event is used, a sedimentation rate of ~21 to 52 cm/kyr (without considering compaction) for the T-OAE NCIE interval in the Da'anzhai Member in the DCD section is obtained. This sedimentation rate is higher than those of approximately 8–19 cm/kyr and 13–30 cm/kyr that would be implied for the Da'anzhai Member in Cores A and B, respectively, as studied by Xu et al. (2017). Estimations of the sedimentation rates in the Da'anzhai Member based on cyclostratigraphy analysis yielded averages of 7.2–9.1 cm/kyr (Qiu et al., 2023). The higher sedimentation rate may also have led to a broader NCIE in the Sichuan Basin compared to the sharper T-OAE NCIE in the Anya section of the Ordos Basin, where the T-OAE NCIE interval is only ~13 m thick (e.g., Jin et al., 2022), implying a sedimentation rate between and ~3 to 8 cm/kyr. These differences may be due to local variability of sedimentation rates between basins and within the Sichuan Basin during the Toarcian time.

It should be noted that the $\delta^{13}\text{C}_{\text{carb}}$ values from ~98 to ~104 m exhibit a negative trend in the bulk limestones, whereas the $\delta^{13}\text{C}_{\text{org}}$ values show the opposite trend, indicating a decoupling of the carbon isotope signals. This decoupling may reflect the differences in the origin and composition of the bulk carbonate fraction, as $\delta^{13}\text{C}_{\text{carb}}$ values of bulk limestones represents a mixed signal from different carbonate components, or the input of $\delta^{13}\text{C}$ -depleted carbon into the water column. The facies change from deeper (limestones) to shallower (sandstone-dominated Lianggaoshan Formation) environments suggests a lake-level decrease, which could have enhanced bottom-water oxygenation. Increased bottom water oxygen levels make organic matter more susceptible to oxidation (e.g., Li et al., 2023), potentially releasing ^{12}C -enriched carbon into the water column. This lighter carbon could then be incorporated into coeval inorganic carbon pool (e.g., carbonate sediments), resulting in the negative $\delta^{13}\text{C}_{\text{carb}}$ values.

5.5. The terrestrial record of the T-OAE NCIE in China: A review

The $\delta^{13}\text{C}_{\text{org}}$ values in lacustrine sediments represent the carbon isotope composition of a mixed kerogen source including terrestrial (plant debris, barks fragments, leaves, and plant waxes) and aquatic (from algae or bacterial) organic matter (e.g., Diefendorf and Freimuth, 2017). As such, the exact composition of the organic matter, especially the contribution of different kerogen types (i.e., Type I, II, and III kerogen), can significantly affect the $\delta^{13}\text{C}_{\text{org}}$ values of bulk rocks (e.g., Jin et al., 2021; Gama and Schwark, 2023). This is because Type I kerogen, mainly derived from lake algal sources, typically has more negative $\delta^{13}\text{C}_{\text{org}}$ values, while Type III kerogen, mainly derived from terrestrial plant organic matter, often has more positive $\delta^{13}\text{C}_{\text{org}}$ values (Meyers, 1994).

The Rock-Eval pyrolysis results for the Da'anzhai Member shales indicate a shift in organic matter source, changing from Type II/III kerogen dominance in the pre-T-OAE NCIE interval to Type I/II kerogen dominance within the T-OAE NCIE interval (Xu et al., 2021). This suggests an increased contribution of aquatic algae-derived organic matter during the T-OAE interval (Xu et al., 2021). Additionally, another study revealed that the organic matter source in the middle part of the Da'anzhai Member (within the T-OAE NCIE interval) is composed of Type II and Type III kerogen, indicating a mixed organic matter origin from both aquatic and terrigenous sources (Wang et al., 2024). Based on the classification framework of Tyson (1995) and our palynofacies data, Type II kerogen is also identified as the predomi-

nant type of sedimentary organic matter in the DCD section (Figs. 4 and 5). Even though no clear covariance exists between *Botryococcus* algae and $\delta^{13}\text{C}_{\text{org}}$ ($r^2 = 0.1$, $n = 28$, Supplementary File Fig. S4), samples with the highest freshwater algae content are coeval with significantly lighter $\delta^{13}\text{C}_{\text{org}}$ values ($< -28\text{‰}$). However, the counted *Botryococcus* abundance might underestimate the true representation of this algal group in the DCD section as its morphology and preservation potential is greatly influenced by taphonomic processes (Kumar et al., 2017). Although generally relatively resistant to degradation (Haselwander and Oboh-Ikuenobe, 2015), microbial degradation can result in the loss of various morphological structures of *Botryococcus* leading to amorphous appearance (Kumar et al., 2017) that can limit identification in palynofacies analysis. Notably, similarly low covariance exists between translucent plant tissues such as leaf fragments and cuticles ($r^2 = 0.2$, $n = 28$, Supplementary File Fig. S4) but there is a higher negative correlation between opaque black-brown wood and $\delta^{13}\text{C}_{\text{org}}$ ($r^2 = 0.6$, $n = 28$, Supplementary File Fig. S4). Xu et al. (2017) showed clear correlation between high pyrolysis HI values and more negative $\delta^{13}\text{C}_{\text{org}}$ values, however, the lack of strong covariance between palynofacies groups and organic carbon isotope ratios in the DCD section might indicate that the $\delta^{13}\text{C}_{\text{org}}$ values here are derived from the mixing of different plant materials or kerogen types each with its own isotopic signature.

Opaque phytoclasts such as black-brown wood particles are thought to be formed by desiccation, oxidation, and fungal degradation of woody material in aerobic conditions in the upper part of soils and peats often linked to variable climate with dry periods and water table fluctuations (e.g., Tyson, 1995; Rodrigues et al., 2019, 2021). However, the abundance of this palynofacies group markedly decreased during the main phase of the NCIE in the Sichuan Basin, with increases in non-opaque phytoclasts (biostructured wood and translucent plant tissues, Fig. 4) coeval with more negative $\delta^{13}\text{C}_{\text{org}}$ values. This rise in the abundance of non-opaque phytoclasts was likely a global trend across the T-OAE NCIE (e.g., Baranyi et al., 2016; Pieńkowski et al., 2016; Slater et al., 2019; Jin et al., 2020; Galasso et al., 2021; Rodrigues et al., 2021), and can be linked to enhanced hydrological cycling that resulted in increased fluvial discharge into lacustrine/marine basins and shortening of the transport pathway (and transport time) for organic particles thus limiting the potential for degradation.

The $\delta^{13}\text{C}_{\text{org}}$ values in Lower Jurassic sediments and the amplitudes of the T-OAE NCIEs from Chinese lacustrine basins formed at different latitudes exhibit significant differences (Figs. 8 and 9). The $\delta^{13}\text{C}_{\text{org}}$ values of lacustrine sediments from higher latitude regions (such as the Jinyang or Tarim basins) are higher compared to those from the lower latitude Sichuan Basin (except for the data from Core B of Xu et al., 2017; see in Fig. 9). Changes in the type of kerogen and the contribution of aquatic algal and/or bacterial organic matter can exert a control on the organic carbon isotope ratios (e.g., Meyers, 1994). In the Sichuan Basin, the organic matter changed from kerogen Type II/III in the pre-T-OAE NCIE interval to Type I/II during the NCIE (Xu et al., 2021). This is also confirmed here by the higher *Botryococcus* and/or AOM abundances at several horizons in the DCD section (Fig. 4). Meanwhile, both in the Ordos and Jinyang basins the kerogen remained mainly Type II and III during the T-OAE (Liu et al., 2020b; Li et al., 2023).

The Sichuan Basin data (Xu et al., 2017, 2021, and this study) indicate the presence of marine acritarchs (Fig. 4). Since a potential marine connection is feasible for the Sichuan Basin during the Toarcian interval (Liu et al., 2024), a contribution from marine organic matter cannot be excluded. As such, the variable content of marine organic matter could also account for the observed differences in $\delta^{13}\text{C}_{\text{org}}$ values across different basins in China (Fig. 9).

Further differences in the $\delta^{13}\text{C}_{\text{org}}$ values can be expected from the varying climate belts and vegetation patterns across a latitudinal gradient in China. At present, C_3 plant wax $\delta^{13}\text{C}$ variations show

distinct latitudinal trends linked to MAP (mean annual precipitation) and MAT (mean annual temperature) changes (Kohn, 2010; Liu and An, 2020). However, modern and Jurassic landscapes, vegetation patterns, and hydroclimate are not entirely comparable. For example, several plant groups, e.g., seed ferns, Bennettitales or cheirolepid (which were key elements of the Jurassic floras) are extinct today. Also, at present MAP is generally decreasing from the tropics towards higher latitudes (Liu and An, 2020), but paleontological and sedimentological evidence implies that the climate in the Early Jurassic of China was characterized by a gradient with more temperate and humid conditions towards the north and the Panthalassa coastline (e.g., Deng et al., 2017). The Jinyang basin was located in a warm and humid temperate climate with a rich flora of Osmundaceae and Ginkgo-related plants (Liu et al., 2019, 2020b). In contrast, the Tarim, Qaidam, and Ordos basins were located in inland areas in the tropical-subtropical semi-arid and semi-humid temperate with more pronounced seasonality and a flora dominated by cycads, Dipteridaceae, and Matoniaceae ferns (e.g., Deng et al., 2017; Baranyi et al., 2023, 2024). The Sichuan Basin was located closer to the tropical-subtropical arid climate province (e.g., Deng et al., 2017; Xu et al., 2017). It is evident that climate, humidity, and plant types underwent changes during the Toarcian (Deng et al., 2017; Baranyi et al., 2024). In the Qaidam, Ordos and Sichuan basins the diversity of the plant community significantly decreased, and was associated with the proliferation of Cheirolepidiaceae and enhanced seasonality during the T-OAE NCIE (e.g., Wang et al., 2005; Deng et al., 2017; Jin et al., 2022; Baranyi et al., 2024). At higher latitudes or closer to the Panthalassa shoreline, vegetation remained more constant with mesophytic and hygrophytic plants (e.g., ferns, cycads, and seed ferns) (Deng et al., 2017; Zhang et al., 2022; Baranyi et al., 2024). It is thus conceivable that differences in the $\delta^{13}\text{C}_{\text{org}}$ values observed in Chinese lacustrine basins (Fig. 9) could be linked to vegetation changes influenced by latitudinal effects. These shifts may have, in turn, contributed to the differences in the amplitudes of the T-OAE NCIE in different sections and cores (Fig. 8).

5.6. Comparison of Early Jurassic carbon cycle perturbations from terrestrial and marine records

Franceschi et al. (2023) presented a composite carbon isotope record spanning the Zhenzhuchong, Dongyuemiao, Ma'anshan members, and the lowest part of the Da'anzhai member in the study area, covering the Sinemurian to the earliest Toarcian (Early Jurassic). Combined with the palynological and $\delta^{13}\text{C}_{\text{org}}$ data of Franceschi et al. (2023), we reconstructed a long-term carbon isotope record spanning the Sinemurian to Toarcian. We then compared this long-time record with that from the marine Mochras Farm (Llanbedr) Borehole in Wales, UK (Storm et al., 2020; Fig. 10). Franceschi et al. (2023) reported several carbon isotope perturbations from the Sinemurian to Pliensbachian. The first, an abrupt ca. -8‰ negative shift, named n-CIE 1, occurred at the boundary between the Sinemurian and Pliensbachian stages and was correlated to the Sinemurian–Pliensbachian Boundary Event, which is well-documented in the Tethyan region (Peti et al., 2017; Schöhlhorn et al., 2020; Storm et al., 2020, and reference therein). Subsequently, the sharp n-CIE 2 in the Franceschi et al. (2023) is correlated to an interval that Storm et al. (2020) recognized (Fig. 10). This NCIE is considered equivalent to a cluster of carbon isotope excursions observed during the 'Davoei-Stokesi negative event,' the 'Stokesi positive event,' and the 'Subnodusus negative event' identified by (Peti et al., 2017) in the Sancerre-Couy core in the Paris Basin (Fig. 10). The n-CIE 3 of Franceschi et al. (2023) can be correlated to an NCIE that is observed at the boundary between *margaritatus* and *spinatum* ammonite zones at Mochras (e.g., Storm et al., 2020; 'Spinatum NCIE' in Fig. 10). The n-CIE 4 in the composite section, identified by Franceschi et al. (2023),

is also observed in the same stratigraphic interval in the DCD section, which we identify as the Pl-To NCIE (Figs. 2 and 8). The Pl-To NCIE has also been reported in the lacustrine sediments from the Ordos Basin (Jin et al., 2020) and also plausibly in the Jinyang Basin (Liu et al., 2020b) (Fig. 8). As mentioned before, our correlation of the T-OAE, implies sedimentation rates in the Toarcian portion of the DCD section ranging between 21 to 52 cm/kyr if the age model proposed by Kemp et al. (2024) for the T-OAE is considered. The correlation proposed in Fig. 10, therefore, evidences a very large change in sedimentation rate between the Pliensbachian and the Toarcian in the DCD record. Given that the Pliensbachian lasted approximately 8.7 million years, the corresponding part of the DCD would be characterized by a sedimentation rate of approximately 0.7 cm/kyr, which is much lower than that in the Toarcian. Such a difference might be explained by the existence of yet unrecognized hiatuses in the DCD section. Refining age constraints and developing more robust age models for the DCD may help explain this phenomenon and improve future correlations.

Despite these uncertainties, that call for further investigations, our identification of key carbon isotope events in our terrestrial record and their correlation with the marine domain strongly supports the inference that these events were globally significant. However, the causes of the carbon cycle perturbations that preceded the T-OAE are not yet well established. Large-scale volcanic eruptions are often considered direct or indirect triggers for carbon cycle perturbations, and it is noteworthy perhaps that these Early Jurassic carbon cycle events coincided with increased tectonic and volcanic activity related to late-phase Central Atlantic Magmatic Province (CAMP) or magmatism linked to the rifting that brought to the breakup of Pangea (e.g., Ruhl et al., 2016, 2022; Jiang et al., 2025). In addition to that, local factors, such as an enhanced lake methane cycling, could have been important contributors to Early Jurassic environmental perturbations e.g., by amplifying the magnitude of the T-OAE NCIE (Fig. 8; Huang et al., 2024), although the full extent of its influence on the organic carbon isotopes composition of lake sediments is not fully understood yet.

6. Conclusions

This paper presents high-resolution carbon and oxygen isotope data, along with palynological data, from the Sichuan lacustrine sediments of the Dacao 'D' section. On the basis of the palynological assemblages, the Dacao 'D' section is dated to late Pliensbachian–Toarcian in age. Petrological study of the lacustrine carbonates (bulk limestone and biogenic shells) suggests that they were minimally or negligibly influenced by diagenesis, and thus preserve the original carbon isotope composition of the Sichuan paleolake. A -2.9‰ negative carbon isotope excursion (NCIE) occurs coincident with Pliensbachian–Toarcian boundary, and a more pronounced NCIE is recorded above, which we attribute to the globally recognized Toarcian Oceanic Anoxic Event (T-OAE) NCIE. Detailed analysis of the carbonate carbon and oxygen isotope values indicates that the Sichuan paleolake operated as a semi-closed system before the T-OAE NCIE, transitioning to a relatively open system during the event.

Carbon isotope data from the Sichuan Basin have lower $\delta^{13}\text{C}$ values compared to basins at higher latitudes, which is tentatively attributed to the influence of vegetation changes across latitudinal gradients (i.e., linked to hydroclimate) and differing contributions of aquatic algal and marine organic matter. These distinct $\delta^{13}\text{C}$ signatures across terrestrial basins in China may be a reason why the T-OAE NCIE identified in different basins has varying magnitude. Additionally, enhanced methane cycling in the lake sediments may have also played an important role in influencing carbon isotope values and thus the magnitudes of NCIEs during this time interval.

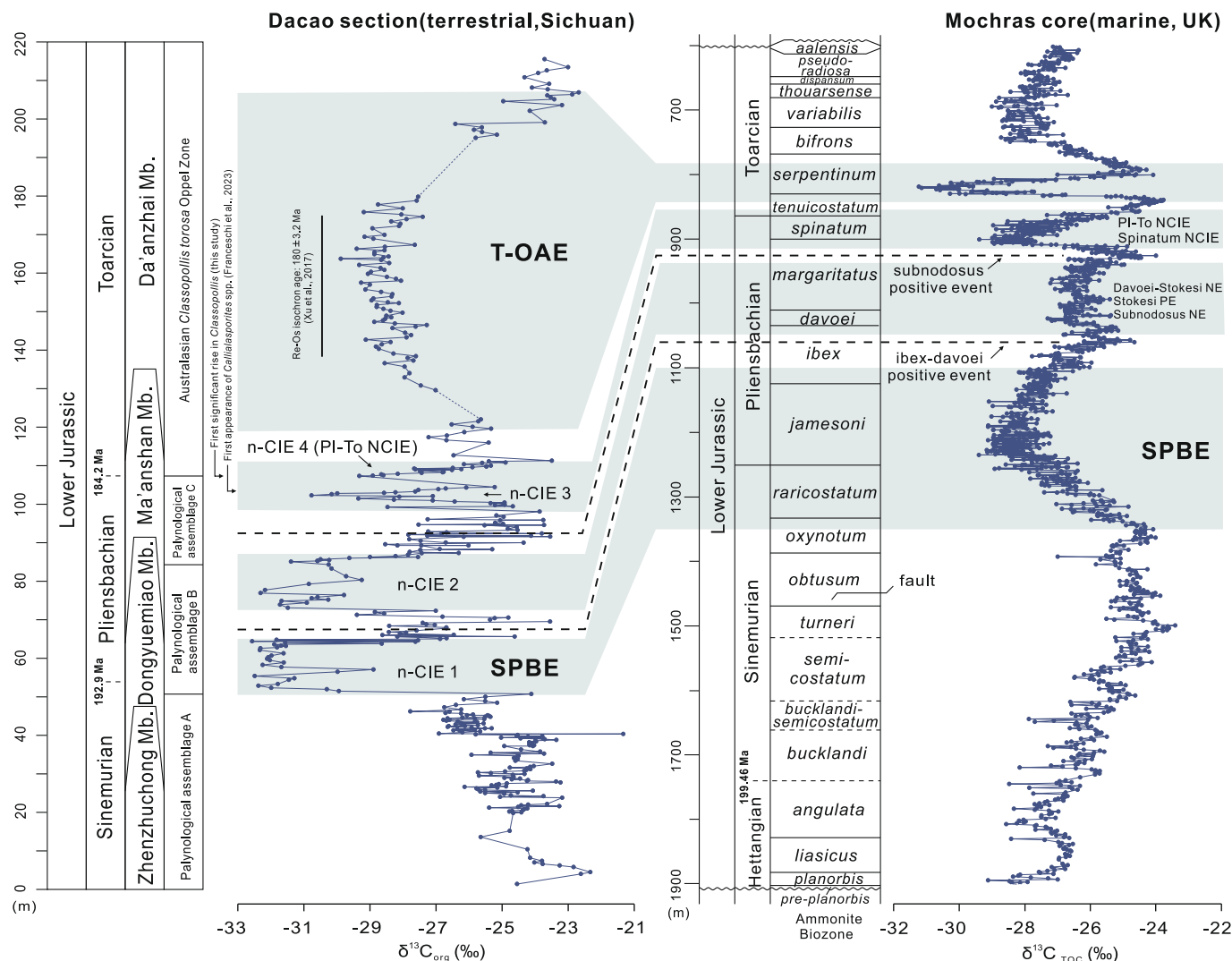


Fig. 10. Comparison of the organic carbon-isotope records spanning Pliensbachian–Toarcian stages between Dacao section (terrestrial) and Mochras core (marine). The isotope curve from Dacao section is composited, i.e., the isotope values of Zhenzhuchong, Dongyuemiao, Ma'anshan members are from Franceschi et al. (2023), while the isotope values of Da'anzhai Member are from this study. The data from the Mochras core are sourced from Storm et al. (2020). SPBE: Sinemurian–Pliensbachian Boundary Event. Palynological assemblage A is dominated by *Chasmatosporites* spp., *Cycadopites* spp., *Marsupipollenites* sp., *Cuneatisporites* sp., *Alisporites* sp., *Auritulinasporites*, *Deltoidospora* sp.; Palynological assemblage B is dominated by *Deltoidospora* spp., *Apiculatisporites* spp., *Concavisporites* spp., *Lycopodiacidites* spp., *Asseretospora amplexiformis*, *Auritulinasporites* sp., *Matonisporites crassiangulatus*, cf. *Cicatricosisporites* sp., *Cibotiumspora* sp.; Palynological assemblage C is dominated by *Chasmatosporites* spp., *Cuneatisporites* spp., *Classopollis* spp., *Sciadopityspollenites* spp. (*Cerebropollenites*), *Callialasporites* spp., *Deltoidospora* sp. (Franceschi et al., 2023). The dates for the boundaries of the Hettangian, Sinemurian, Pliensbachian, and Toarcian stages are based on Gradstein et al. (2020). It should be noted that the carbon isotope correlation between the Dacao section and the Mochras core does not take potential hiatuses in the Dacao section into account.

Combining our new data with the previous results of Franceschi et al. (2023), a long-term carbon isotope curve spanning from the Sinemurian to the Toarcian stages in Sichuan terrestrial rocks has been reconstructed, which can be correlated with similarly long marine records. As such, our compilation of lacustrine carbon isotope data from the terrestrial biosphere enhances our understanding of the temporal and spatial variations in carbon cycling during the Early Jurassic.

Declaration of Generative AI and AI-assisted technologies in the writing process

During the preparation of this work the author(s) used Chat GPT in order to improve the language of the manuscript. After using this tool/service, the author(s) reviewed and edited the content as needed and take(s) full responsibility for the content of the publication.

CRediT authorship contribution statement

Xin Jin: Writing – original draft, Visualization, Supervision, Methodology, Investigation, Funding acquisition, Formal analysis, Conceptualization. **Viktória Baranyi:** Writing – original draft, Visualization, Methodology, Formal analysis. **David B. Kemp:** Writing – review & editing. **Zhiqiang Shi:** Writing – review & editing. **Hao Zou:** Writing – review & editing. **Binbing Li:** Writing – review & editing, Visualization. **Yunwang Zhang:** Writing – review & editing. **Marco Franceschi:** Writing – review & editing, Supervision, Investigation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.gsf.2025.102073>.

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