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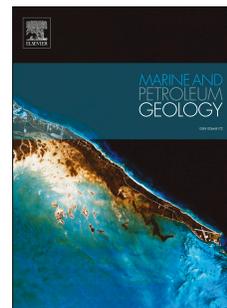
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Journal Pre-proof

1 ***Carboniferous–Permian transgression/regression mechanisms in the Eastern***  
2 ***Ordos Basin and their sea-level spatiotemporal variability: Insights from***  
3 ***source-to-sink systems***

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9 **Abstract:** Developed on the North China Craton, the intracratonic Ordos Basin contains a  
10 complete Paleozoic-to-Cenozoic sediment record, allowing for long-term paleoenvironmental and  
11 climate change investigation. During the Carboniferous–Permian period, convergence between  
12 the North China block and the paleo-Yangtze plate to the south resulted in a general marine  
13 regression characterized by a series of second-order transgression/regression cycles  
14 diachronous along the eastern margin of the Ordos. However, the detailed mechanisms that  
15 induced these cycles, as well as the associated paleoecological changes, remain unknown. In  
16 this study, we integrate multiple indices, including  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ , rare earth element (REE),  
17 paleontological assemblages, and clay content ( $w(\text{chlorite} + \text{kaolinite})/w(\text{Illite})$  ratio) planar  
18 distribution, to restore the paleo-water depth distribution. These parameters are then used to  
19 reconstruct the paleo-sea level from the Pennsylvanian to the middle Permian. We conclude that  
20 the direction of second-order transgression/regression was mainly toward the east during the  
21 Pennsylvanian–early Permian and switched clockwise toward the north during the middle  
22 Permian. We suggest that the second-order cycles, diachronous in space and time, are mainly

23 linked to local variations in sediment supply and regional uplift. Using detrital zircon U–Pb data  
24 and the REE and trace element content and heavy mineral assemblages (HMA), we estimate the  
25 sediment provenance area. The sediment volumes deposited in the basin through time are  
26 obtained using 2D seismic data. During the Carboniferous, the coarse-grained sediments  
27 deposited in the eastern Ordos were derived from the uplifting Helan Mountain (Qiandam–Qilian  
28 orogenic belt). By the middle Permian, the detrital material became multi-sourced, thus issuing  
29 the Yinshan range to the north and the Qinling range to the south. We then integrate the  
30 description of numerous core samples with electric log and 2D seismic data to reconstruct the  
31 sediment facies associations across the first-order regression from the Carboniferous tidal flat  
32 depositional system to the middle Permian prograding fluvial delta system. According to the  
33 transfer of the glacial epoch, the sedimentation rate, and the transgression/regression rate above,  
34 we classify the evolution process into three patterns: low-transgression rate and less-sediment  
35 supply pattern (the late Carboniferous), high regression rate and mass sediment supply pattern  
36 (the north block during the early Permian), and low regression rate and mass sediment supply  
37 pattern (the south block during the Sakmarian stage). Lastly, with the quantitative calculation of  
38 the source-to-sink (S2S) parameters, including the S2S system volume and the elevated height of  
39 regional uplift with the 2D seismic data, we propose the mechanism of transgression/ regression  
40 in the Ordos Basin, responding to the above three pattern s. The first pattern was controlled by  
41 regional uplift, whereas the second pattern was controlled by sediment supply. As to the third one,  
42 uplift and sediment supply could affect the transgression/regression process.

43       Keywords : Transgression mechanism, Provenance identification, Detrital geochronology data,  
44 Paleontology data, Pennsylvanian–Permian transition

## 45 1. Introduction

46 An intracratonic basins is characterized by the preponderance of thermal subsidence over  
47 tectonic subsidence, making for largely stable long-term subsidence rates. Sedimentary  
48 sequences in this type of basin generally preserve a complete sedimentary record that allows  
49 deciphering the large-scale paleoenvironmental evolution (*Sloss, 1963; McLaughlin et al., 2004;*  
50 *Zhu et al., 2008; Hoffmann et al., 2009*). When affected by marine incursions, these basins  
51 provide a good record of water-depth fluctuations and associated transgression/regression  
52 mechanisms (*Nikishin et al., 1996; Martins-Neto et al., 2001; Alkmim and Martins-Neto, 2012;*  
53 *Bumby et al., 2012*). As a typical intracratonic basin in the western part of the North China Block,  
54 the Ordos Basin (*Wang et al., 2019; Yin et al., 2019; Shi et al., 2020*) contains a complete  
55 Carboniferous–Permian sedimentary record of transgression/regression cycles closely related in  
56 time to a transition from a glacial period to an interglacial one (*Tabor and Poulsen, 2008; Koch*  
57 *and Frank, 2011; Xu et al., 2020*).

58 Previous studies have pointed out that with the end of the Late Paleozoic Ice Age (LPIA)  
59 during the late Permian, the increase in rainfall resulted in additional sediment supply worldwide,  
60 leading to a marine regression process in shallow-marine basins (*Golonka and Ford, 2000; Tabor*  
61 *and Poulsen, 2008; Koch and Frank, 2011*). However, *Stollhofen et al., (2000)* pointed out that  
62 the transgression/regression process shows a spatiotemporal variability within the large study  
63 area. Using stable isotope analysis ( $\delta^{13}\text{C}$  and  $\delta^{15}\text{N}$ ) in the eastern margin area (EMA) of the  
64 Ordos Basin, *Xu et al. (2020)* further highlighted the complexity of the transgression/regression  
65 events. Meanwhile, *Zhu et al. (2008) and Zhang et al. (1997)* suggested that during the  
66 Carboniferous–Permian, the water depth in the EMA continuously declined due to uplift of the

67 basin margins (**FIG. 1**). Finally, Jiang and Wang (2012) demonstrated that the main direction of  
68 the transgression/regression phases changed constantly through time, adding complexity to the  
69 water-depth fluctuation pattern.

70 Several authors have suggested that tectonic movements linked to continental accretions  
71 along the margins of the North China Craton had been the main driving mechanism of water  
72 depth changes in the basin (*Qin et al., 2002; Yang et al., 2005*). However, by focusing their  
73 models on the effects of tectonics, these studies neglected the influence of changes in climate  
74 and sediment supply rates, which are also factors of water-depth fluctuation, particularly in  
75 shallow, marginal basins (*Haq et al., 1987; Haq and Schutter, 2008a; Labaune et al., 2008; De*  
76 *Falco et al., 2015; Fu et al., 2019*). Indeed, existing studies have shown that on the eastern  
77 margin of the Ordos Basin, the direction of the main sediment supply changed during the  
78 Carboniferous–Permian period, implying changes in regional topography and erosion pattern  
79 (*Kröner et al., 2006; Darby and Gehrels, 2006; Milne et al. 2009; Zhu et al., 2011*). Regarding  
80 intracratonic basins, *Klein (1982)* suggested that an arid climate can potentially change the ratio  
81 between clastic and carbonate deposits and indirectly affect water depth.

82 In this study, we point out the water-depth variations and derive local and regional  
83 transgression/regression cycles for the Pennsylvanian-to-early-Permian period. We then integrate  
84 detrital zircon U-Pb analysis, heavy mineral assemblages (HMA), and trace elements indexes to  
85 determine major changes in sediment sources, climate, transport conditions, and depositional  
86 environments. We then quantitatively compare the sediment composition and supply rates to  
87 water-depth fluctuations and propose three major source-to-sink (S2S) patterns for the late  
88 Paleozoic eastern Ordos Basin: low-transgression rate and limited sediment supply (LT–LS),

89 low-transgression rate and massive sediment supply (LT–MS), and high-transgression rate and  
90 massive sediment supply (HT–MS). Finally, the response to these various patterns in terms of  
91 water-depth variations is described to document the transgression/regression cycles during the  
92 final stage of the LPIA.

## 93 **2.Geological Setting**

94 The Mesozoic-to-Cenozoic geologic history of Central China has been strongly influenced by  
95 continental accretion events along the southern margin of the Eurasian continent. These  
96 geodynamic events drove the formation and sometimes the reactivation of mountain ranges such  
97 as the Lvliang and Helen-Liupan Mountains east and west of the Ordos Basin, the Yinshan  
98 Mountain to the north and the Qinling Mountain to the south (**FIG. 1**). The Ordos Basin itself is  
99 formed by several depressions such as the East margin, Hetao, Weihe, and Yichuan depressions  
100 while major faults control the basin boundaries. The evolution of the Ordos Basin can be divided  
101 into five main stages: it initiated as a Cambrian–Early Ordovician continental rim basin, evolving  
102 towards an intracratonic basin in a Middle Ordovician–Middle Triassic generally compressive  
103 geodynamic setting; during the Late Triassic–Early Cretaceous the basin was affected by thermal  
104 subsidence before being inverted and eroded during the late Cretaceous to Neogene  
105 compressive event (only some Cenozoic loess are preserved in the south part the basin (*Yuan et*  
106 *al., 2007*)).

107 The sedimentary sequence preserved in the EMA ranges from 7 km to 10 km in thickness  
108 (*Yang et al., 2005*). During the Ordovician to Mississippian period, tectonic activity induced uplift of  
109 the eastern margin of the Ordos Basin (present-day reference frame) leading to a regional  
110 parallel unconformity between the basement and the Pennsylvanian deposits (*Yang et al., 2005*).

111 The Pennsylvanian to Permian series are divided into the following formations: Penci (C<sub>2</sub>b),  
112 Taiyuan (C<sub>2</sub>t), Shanxi (P<sub>1</sub>s itself divided into the upper and lower P<sub>1</sub>s members), Lower Shihotse  
113 (P<sub>2</sub>sh), Upper Shihotse (P<sub>3</sub>sh), and Shiqianfeng (P<sub>3</sub>s) (**FIG.2**).

114 During the deposition of the Penci Formation (Fm.), the sedimentary record displays a sharp  
115 regression phase generally associated to the uplift of the central part of the Ordos region. The  
116 facies association evolves from a shallow marine setting to a coastal swamp environment. Based  
117 on core sample interpretation, *Zhang et al. (1997)* and *Jiang and Wang (2012)* indicated that  
118 alluvial fan or braided delta depositional environments dominated in the northern part of the study  
119 area. *Stanley and Powell (2003)* stated that in the late Pennsylvanian period, the arid climate and  
120 the correlatively reduced amount of sediment injected into the basin resulted in the development  
121 of thick limestone and coal seams in the Taiyuan Fm. However, the sea level showed a gentle  
122 transgression, and a carbonate platform dominated during this period.

123 In the early Permian, the convergence between the South China and North China plates  
124 largely modified the paleomorphology and paleoecology of the Ordos region. Furthermore, the  
125 increasing sediment supply from multiple directions led to a gentle regression from the Asselian  
126 to the Kungurian stages (Shanxi Fm.). The combined studies of *Zhang et al. (1997)*, *Zhu et al.*  
127 *(2008)*, and *Xin et al. (2018)* pointed out that deltaic environments dominated in the EMA, with  
128 large marshes separating the distributary channels. With the decrease in sea level, the  
129 meandering channels in the delta plains were replaced by braided river environments marked by  
130 sand bars and crevasse splays.

### 131 **3.Data and methodology**

#### 132 **3.1 Research data**

133 The EMA of the Ordos Basin has been explored by 3700 km of 2D seismic and 1000 km<sup>2</sup> of  
134 3D high-resolution seismic data. Furthermore, we get access to 15 wells drilled in the EMA. The  
135 positions of the cross-well seismic sections used in this study are shown in **FIG. 1a**. Electric well-  
136 log data, including gamma ray (GR), were acquired and processed by Schlumberger®. We get  
137 the core samples from wells DJ12, Y253, DJ43, G10, G6, and DJ19 (**FIG. 3**). Heavy and clay  
138 mineral content and their scanning electron microscope (SEM) photographs as well as whole rock  
139 trace elements analysis from wells H4, Y787, G3, DJ43, DJ53, Y521, Y653, and Y432 were  
140 processed by the Geological Laboratory of the Research Institute of C.M.B. company (China  
141 National Petroleum Corporation CNPC®). The paleontological identifications, including pollen and  
142 plant assemblages from wells G3, H4, DJ43, DJ53, and Y787, were completed by the Langfang  
143 Branch of Petrochina Exploration and Development Research Institute (CNPC®).

144 LA-ICP-MS U–Pb analysis of detrital zircons from four core samples (DJ53 1942.23 m, H5  
145 1968.1 m, DJ51 2264.85 m, and G2 2048.1 m) was completed by the Chinese Academy of  
146 Geological Sciences® (**Appendix TABLES 1–4**). Laser ablation was conducted using a  
147 NWR193UC 193 nm deep ultraviolet laser ablation sampling system (Elemental Scientific Lasers,  
148 USA). Mass spectrometry was calculated using an Agilent 7900 inductively coupled plasma mass  
149 spectrometer (Agilent, USA). One hundred twenty detrital zircons from each sample were  
150 analyzed with a spot size of 30 µm. Every five analyses were followed by one analysis of  
151 standard zircon 91500 and GJ. Common lead correction was conducted using  
152 ComPbCorr#3\_15G. The U–Pb age calculations and Concordia diagram were completed using  
153 *IsoplotR*® software provided by University College London, UK ([http://pieter-  
154 vermeesch.es.ucl.ac.uk/shiny/IsoplotR/](http://pieter-vermeesch.es.ucl.ac.uk/shiny/IsoplotR/)). Only zircon ages concordant within 90%–110% were

155 considered in the subsequent discussion.

156  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  data were analyzed from whole-sediment mudstone, limestone, or coal  
157 samples from cores Y787, DJ53, DJ43, H4, and G4 (some carbonate poor layers did not provide  
158 reliable data and results were not reported). The C and O-isotope analyses results are  
159 normalized to the Pee Dee Belemnite (PDB) standards with a precision of  $\pm 0.1\text{‰}$  for  $\delta^{13}\text{C}$  and  
160  $\pm 0.2\text{‰}$  for  $\delta^{18}\text{O}$ .

### 161 **3.2 Methodology**

162 1) Sedimentology and stratigraphy analysis from drill cores and seismic data

163 Facies association were established from core samples according to the sedimentary  
164 structure (**FIG.3**) and the grain-size distribution (**FIG.4**). Geophysical log data were then used to  
165 constrain the depositional environment setting, such as the bell or cylinder shape of the GR curve  
166 (indicating a decreasing flow rate), or a sharp decrease of the DEN curve (or sonic/neutron  
167 porosity curve) indicating the occurrence of coal. Finally, associating well and 2D seismic data we  
168 reconstructed the stratigraphic and depositional environment pattern of the EMA of the Ordos  
169 Basin through time. The 2D seismic profiles were studied and interpreted following the  
170 methodology previously established by *Pu et al. (2009)*, *Xu et al. (2016)*, and *Anees et al. (2019)*  
171 in the Ordos Basin. Three types of seismic facies were used as guidelines: the concave, convex,  
172 and plate shapes. Following that approach, a seismic package with a concave boundary was  
173 generally interpreted as an eroding channel. Associating the log data to the seismic image, it was  
174 possible to distinguish between deposited tidal channels, braided channel belts and depositional  
175 meandering channel belts. Convex-shaped seismic facies were generally interpreted as  
176 depositional tidal barriers. The plate-shaped seismic facies, characterized by a high amplitude

177 seismic reflection, generally indicate the development of carbonated tidal platforms.

178 2) Identification of sediment provenance area and calculation of supply rates.

179 By combining the results of previous studies constraining the provenance area of the EMA  
180 during the late Paleozoic (*Darby and Gehrels, 2006; Kröner et al., 2006; Zhu et al., 2011; Song et*  
181 *al., 2012; Xie and Heller, 2013; Sun et al., 2014*), we established a mean zircon U–Pb age  
182 distribution for the EMA. Using both the age data for the basement rocks, the zircon–tourmaline–  
183 rutile (ZTR) index (provides indications on the distance of sediment transportation), grains size,  
184 sediments Rare Earth Elements patterns and informations on the tectonic background (*Darby and*  
185 *Gehrels, 2006; Xie and Heller, 2013*), we evaluated the evolution of the main provenance area  
186 during the interval from Pennsylvanian to the early Permian.

187 The sediment supply rates through time were calculated using the provenance data obtained  
188 from the method described above and a series of selected 2D seismic profiles. The thickness of  
189 the main sediment layers (A parameter hereafter) corresponding to the Bashkirian to Sakmarian  
190 stages was obtained from seismic data. As shale deposits have a much higher compaction  
191 coefficient than sandstone layers we divided the strata using the sand content (over 80%)  
192 observed within the corresponding reference wells (*Yang et al., 2008; Zhang et al., 2009; Zou et*  
193 *al., 2013*). Pre-compaction thicknesses were then calculated using compaction coefficients of  
194  $0.0004 \text{ m}^{-1}$  for mudstone and  $0.00025 \text{ m}^{-1}$  for sandstone (*Zhang et al., 2009*). Finally, the  
195 stratigraphic and geochronology data obtained in reference wells were used as time constrains (T)  
196 to derive sedimentation rates ( $v_{\text{sediment}} = A/T$ ).

197 3) Regional paleo-sea level reconstruction via multi-index integration

198 As sea level fluctuates all the time, the lateral facies variations observed in the sedimentary

199 record may lead to some illusion and contradiction in the reconstructed regional paleo-sea level.  
200 In this study we computed the relative sea level of the Ordos Basin EMA based on the  $\delta^{18}\text{O}$  and  
201  $\delta^{13}\text{C}$  data from the high organic matters mudstone and comparing to the similar curve obtained  
202 by Xin et al. (2018). Our interpretation takes into account the fact that the warming climate during  
203 the Carboniferous–Permian transgression accounted for an increase of  $0.008\text{--}0.01\text{‰m}^{-1}$  of the  
204  $\delta^{18}\text{O}_{(\text{V-PDB})}$  value (Adkins et al., 2002; Elderfield et al., 2012).

205 A second independent model was derived from paleontological data. *Paripters spp.–Linopteris*  
206 *spp.–Conchophyllum* plant association are characteristic of the Bashkirian–Kasimovian stage.  
207 During the Gzhelian–Asselian stage, the representative paleontological assemblage is composed  
208 of *Neurpteris pseudovata* and *Lepidodendron posthumii*. Then, with the warming climate,  
209 *Emplectopteridium alatum–Cathaysiopteris whitei–Emplectopteris triangularis* occurred during the  
210 Sakmarian–Kungurian stage (Hilton and Cleal, 2007). Some species (*Zygopteris* and *Ankyropteris*  
211 *corrugated*) accumulation, including their macro-remains and pollen, reflected an onshore setting  
212 (Phillips and Galtier, 2005; Phillips and Galtier, 2011). Hence, we calculated the ratio of fern  
213 species biomass and the plant fossil identified ( $w_{\text{biomass}}$ ) in G3, H4, DJ43, DJ53, and Y787 and  
214 obtain their responding relative sea level. Furthermore, according to the morphology of all the  
215 identified species in the core samples of the EMA (Hilton and Cleal, 2007; Pfefferkorn and Wang,  
216 2007), broadleaf ferns were prominent during the late Paleozoic although their preservation in the  
217 fossil record was affected by post-deposition sediment reworking processes (Bacon et al., 2016).  
218 Previous studies have pointed out that at the high stand of the sea level, the amount of leaf fossil  
219 in the core samples ( $w_{\text{leaf}}$ ) is larger than the root or stem fossil due to the hydrodynamic process  
220 of tide or wave (Pratt et al., 1978; Gastaldo, 2004; Gee, 2005). Thus, according to the ratio

221 between the leaf fossil and the root/ stem fossil ( $w_{leaf}$ ), we can infer the relative sea level.

222 Clay mineral assemblages provide another evidence for the late Paleozoic relative regional  
223 sea-level proxy. The Panthalassic ocean margin was located east of the EMA and south of the  
224 Paleo–Tethys (<http://www.scotese.com/>). The clay mineral assemblages of the Panthalassic  
225 ocean (*Kemp and Izumi, 2014*) and Paleo–Tethys (*Cheng et al., 2019*) display a high value of  
226 illite. By contrast, the clay mineral assemblage of the potential source area is mainly chlorite and  
227 kaolinite (*Zhang et al., 1998*). Thus, the deposition of the marine original sediment will lead to a  
228 gentle increase of the illite content in the sediment. That is to say, the illite shows a much weaker  
229 positive relationship to the transgression process. According to the method mentioned in *Fu et al.*  
230 (*2019a and b*), we calculated the  $w_{clay} = w(\text{chlorite} + \text{kaolinite}) / w(\text{illite})$ , and variations in this  
231 parameter are inversely correlated to the variations in sea-level.

232 The trace element ratios including  $\Sigma\text{LREE} / \Sigma\text{HREE}$ , Sr/Br, Ce/La, Rb/Zr,  $\text{CaO} / (\text{MgO} \times \text{Al}_2\text{O}_3)$ ,  
233 and  $\text{V} / (\text{V} + \text{Ni})$  were also obtained from core samples in wells G3, H4, DJ43, DJ53, and Y787 .  
234 Values of  $\text{V} / (\text{V} + \text{Ni}) > 0.54$ , Sr/Br > 0.5, and Ce/La > 1.5 are interpreted as characteristic of a  
235 shallow-marine environment, the other values indicating a continental setting (*Rimmer, 2004*).  
236 Referring to the above critical condition, we obtained the relative regional sea-level fluctuation in  
237 accordance with the trace element in.

238 Transgression/regression rate calculation: based on the transgression/regression directions  
239 established above, we used 2D seismic data to identify the successive transgression events and  
240 calculate their distance of propagation (S). As indicated by the sediment thermal evolution history  
241 diagram previously obtained on these series (see *Yu et al., 2012; Ren et al., 2007*), the  
242 subsidence rates varied through time but remained spatially coherent (no large variations

243 throughout the study area at a given time). Subsidence was nearly null during the Pennsylvanian  
 244 interval, while it reached 30 m/Ma during the early Permian. Accordingly, the subsidence rate was  
 245 used as an input parameter in the transgression/regression rate calculationA following the  
 246 equation:

$$247 \quad S = (D_{\text{compaction}} + D_{\text{subsidence}}) / \cos(\alpha),$$

248  $D_{\text{compaction}}$  is the decompacted thickness of mudstone in the strata;  $D_{\text{subsidence}}$  is the  
 249 accommodation of regional subsidence;  $\alpha$  represents the slope angle calculated from the  
 250 interpretation of the seismic profile.

251 Finally, by eliminating the subsidence impact factors, we calculated the  
 252 transgression/regression distance (S) using ( $v_{\text{sea}}$ ) as  $v_{\text{sea}} = S/T$ , in which T is time.

#### 253 **4. Lithofacies identification and association classification**

254 During the transgression/regression process in the Carboniferous–Permian, the grain size and  
 255 lithology exhibited a dramatic change from conglomerate to mudstone. The series are also  
 256 containing well-identified layers of coal, limestone, quartz sandstone, and feldspar sandstone. In  
 257 this study, we defined 6 lithofacies associations using lithofacies previously established by *Zhu et*  
 258 *al. (2008)*, *Yang et al. (2017)*, and *Xin et al. (2018)* in Ordos basin: massive mudstone (**M**);  
 259 bioturbated sandstone (**Sb**); muddy strip composite sandstone (**Scm**); slumped sandstone (**Sd**);  
 260 cross-bedding composite sandstone (**Sc**); cross-bedding sandstone (**Sc**); massive sandstone  
 261 (**Sm**); planar cross-bedded sandstone (**Sp**); massive gravel (**Gm**); planar cross-bedded gravel  
 262 (**Gp**) (**FIG. 3** and **TABLE. 1**).

#### 263 **5. Identification of depositional environments from seismic and well data.**

264 We selected two seismic profiles in an EW direction and calibrated them using wells G6 and

265 DJ12 as references. Based on this calibration, we defined depositional environments and created  
266 a stratigraphic column.

### 267 **5.1 Pennsylvanian depositional environments**

268 As shown from core samples in well G6, the lithofacies association evolved upward from  
269 LFA.5 to LFA.4, LFA.4, LFA.5, and finally LFA.6. Based on lithofacies interpretation in **FIG.3**, the  
270 early Pennsylvanian period was characterized by an initial bidirectional slow-rate traction flow  
271 deposit, evolving upward toward a coarse-grained flow . Finally, facies LFA.5 and LFA.6 indicate  
272 a weak hydrodynamic intensity during the late Pennsylvanian period. The cumulative grain-size  
273 curve in **FIG. 4** also shows this last depositional environment evolution from a mass flow (g and f)  
274 up to a tidal-dominated shallow-water setting (a to e). Additionally, *fusulinid* fossil typical of  
275 shallow marine environments were observed at a depth of 2,431 m in well G6,(*Liu et al., 2015*).  
276 The morphology of the GR curve also indicated the depositional environments in upward layers  
277 as having been transient. In the early Pennsylvanian interval, the well-logging GR curve of  
278 sandstone mainly showed a bell (2455m in G6 **FIG.4**) or cylinder shape (2450 m in G6 **FIG.6**),  
279 indicating a rising sea level and an increase in flow dynamics. For the late Pennsylvanian interval,  
280 the combination of the GR curve and the corresponding density curve (or sonic/neutron porosity  
281 curve) highlight the occurrence of coal layers interbedded in limestone, characterizing a  
282 decreasing sea level and a weakening flow rate. The Pennsylvanian period is thus marked by a  
283 transgression phase followed by regression. However, regionally, the sea level showed a marked  
284 spatiotemporal variability discussed below.

285 Well G6 was used to calibrate the facies pattern on the seismic line shown on **FIG. 4**. The  
286 reflectors showing concave interfaces indicated erosion in tidal channel belts, while convex

287 interfaces are interpreted as sediment accumulation in tidal barriers. The tidal-dominated deposit  
288 (including the barrier and the tidal channel deposit) thus mainly developed during the early  
289 Pennsylvanian period.

## 290 **5.2 Development of depositional environments during the early Permian period**

291 The core samples in well DJ12 show a vertical evolution of lithofacies associations from  
292 LFA.6–LFA.2–LFA.1 (**Table 1**). The high-rate flow deposits of LFA.2 developed during the  
293 Asselian to Sakmarian interval. Deposition of LFA.1 including large mud clasts and lamellar  
294 cross-bedding indicate a slowing flow rate upward. As the cumulative grain-size curve in **FIG. 5**  
295 shows, two curve-pattern segments indicate a decreasing amount of suspended sediments and  
296 a flow-pattern evolution toward a one-directional traction flow. On the corresponding well-logs, the  
297 GR curve developed as a cylinder at the bottom and as a bell shape upward. This again indicate  
298 that the Asselian to Sakmarian interval can be divided into two cycles corresponding to a gentle  
299 transgression process.

300 **FIG 5** displays a series of concave and convex seismic facies according to the seismic  
301 boundaries. Referring to the study of *Zhu et al. (2008)* and *Xin et al. (2018)*, we infer that braided  
302 channels (delta) initially developed, before being replaced upward by meandering channels. The  
303 seismic packages ascribed to the Asselian period show low amplitude and high continuity  
304 patterns. According to *Xin et al. (2018)*, the lateral migration events indicate that sandy  
305 sedimentary facies (such as braided delta) developed in the Asselian period, while upward  
306 toward the Sakmarian interval, we suggest that high amplitude and low continuity events  
307 represent a muddy depositional environment.

## 308 **6. Results**

## 309 6.1 Detrital zircon U–Pb data

310 Detrital zircons U-Pb age data are used to constrain the potential source area of sediments  
311 deposited in the Ordos basin during the Pennsylvanian to the early Permian. Four samples were  
312 selected in the early Permian series from various wells: DJ53 (1942.23 m Asselian), H5 (1968.1  
313 m Asselian), DJ51 (2264.85 m Sakmarian), and G2 (2048.1 m Sakmarian) and representative  
314 zircon CL images are shown on **FIG.6**. For the Pennsylvanian interval we refer to data from *Sun*  
315 *et al. (2014)* (samples JC-A and JC-B). Representative zircon CL images are shown in FIG. 6,  
316 Concordia plots in FIG.7 and individual age spectrums in FIG.8.

317 Samples JC-A and JC-B from the Bashkirian–Gzhelian stage are taken from *Sun et al. (2014)*.  
318 Among the 94 detrital zircons in JC-A and 95 detrital zircons in JC-B, 83 (JC-A) and 94 (JC-B)  
319 have been considered adequate for dating (**FIG. 7**). Cathodoluminescence (CL) images show  
320 oscillatory zoning and Th/U ratios are generally over 0.1, indicating intermediate source rock  
321 types such as granitoids and metamorphic rocks (**FIG.9 a**). The obtained detrital zircon U-Pb  
322 ages can be divided into three groups: 273–307 Ma, 762–974 Ma, and 1168–2543 Ma.

323 In Asselian samples DJ53 and H5, zircon crystals are colorless, mostly elliptical (Fig. 6) and  
324 can be divided into three types according to their external morphology and internal structure. The  
325 first type is rounded/sub-rounded with a length/width ratio of 1:1.5 and shows a well-preserved  
326 magmatic oscillatory zoning. The second type has a length/width ratio of 1:3 and is angular/sub-  
327 angular shaped. The internal structure of zircons from this second group is either homogeneous  
328 or has patchy zoning. This kind of zircon constitutes 20% of each sample, and their age ranges  
329 from 1800 Ma to 1900 Ma (see crystals date at  $1827 \pm 20$  Ma (DJ53) or  $1803 \pm 22$  Ma (H5) on  
330 **FIG. 6**). The third type of zircon is sub-rounded in shape and has a dark magmatic oscillatory

331 zoning implying a metamorphic source, especially for samples over 2000 Ma. A total of 180  
332 crystals were analyzed for those two samples, and 168 of these, with a >90% concordance were  
333 considered adequate (age data in **Appendix Tables 1 and 2**). The individual age are grouped  
334 into four classes: 310–375 Ma, 410–490 Ma, 1650–2300 Ma, and 2300–2500 Ma (**FIG. 7**).

335 Samples DJ51 and G2 were collected from the Sakmarian series. As shown in the CL image  
336 of **FIG. 6**, zircon crystals of DJ51 and G2 are mostly angular/subangular with a length/width ratio  
337 of less than 2:1 and show a dark oscillatory zoning and bright overgrown homogeneous rim. The  
338 Th/U ratio ranging from 0.03 to 0.7 indicates intermediate rock sources. The U–Pb Concordia  
339 plots in **FIG. 5** shows that the concordance of all the grains (85 grains in DJ51 and 94 grains in G2)  
340 is over 90%. The U–Pb age spectrum of DJ51 and G2 reveals two groups at 275 to 450 Ma (peak  
341 value of 375 Ma) and 1750–2000 Ma.

## 342 **6.2 Heavy mineral assemblages**

343 Six types of heavy minerals (rutile, zircon, tourmaline, titanite, apatite and pyrite) were  
344 quantified from core samples and classified in five types using the Heavy Minerals Associations  
345 (HMA) of *Morton and Hallsworth (1994)* and the statistical approach of *Andrew (1985)* : rutile–  
346 dominated (HMA-I), pyrite–dominated (HMA-II), titanite–dominated(HMA-III), tourmaline–  
347 dominated, (HMA-IV), and zircon–dominated (HMA-V) (**FIG. 10**). In order to allow comparison,  
348 the HMA of potential source areas around the Ordos Basin, which includes the Qaidam–Qilian  
349 orogenic belt (QQOB) in the west, Yishan Orogenic belt (YOB) in the north, and Qinling orogenic  
350 belt(QOB) in the south, and their dominated HMA types are obtained from bibliography are listed  
351 in **TABLE.2**.

352 The early Pennsylvanian deposits are characterized by HMA-II and HMA-III associations.

353 Similar associations, dominated by HMA-II are observed in the early Pennsylvanian deposits of  
354 the QQOB, YOB suggesting common sediment sources(*Chen et al., 2009; Du et al., 2017*). The  
355 pyrite content in wells Y521, H4, Y787, and Y653 is over 60%, which suggest that some sediment  
356 may be transported from the QQOB. Besides that, we compared the HMA in DJ43, DJ53, and  
357 that in the YOB (*Ruppen et al., 2014; Li et al., 2014*); both show HMA-III. Based on the increase  
358 in that new source area we speculate that some sediment migrated far south from the northern  
359 Ordos margin.

360 After the deposition of the late Pennsylvanian series, the pyrite content in well Y787 decreased  
361 sharply, whereas the proportion of tourmaline considerably increased. This result suggests that  
362 more sediment from the QQOB were either routed towards another basin or deposited in a  
363 proximal area and not transported into the Ordos sink area. The well in the northern EMA  
364 underwent an increase in zircon and rutile content, suggesting that a substantial amount of  
365 sediment came from the YOB.

366 During the Asselian interval, the most represented heavy minerals are rutile, zircon, tourmaline,  
367 titanite, apatite, and pyrite. The above can be further classified into three types of associations:  
368 HMA-III , HMA-IV, and HMA-V. The average ZTR index in this interval shows an increase from  
369 35% to 73%, indicating that the sediment was deposited in an increasingly distal area, and  
370 detrital material from different provenance areas was mixed.

371 In the Sakmarian stage, the northern EMA showed an increase in titanite content in wells H4,  
372 Y521, Y787, Y653, G3, and Y432, and the average grain size of heavy minerals increased from  
373 0.01–0.15 mm to 0.02–0.03 mm. The average ZTR index shows a decreasing trend, from 52% to  
374 31% that, together with the above parameters suggest a more proximal source, possibly in the

375 QOB affected by uplift at that period (Zhu et al., 2011; Dong et al., 2016).

### 376 **6.3 Trace element and REE patterns**

#### 377 (1) Trace element and REE patterns

378 The chemical stability of La, Ce, Nd, Y, Th, Zr, Hf, Nb, Ti, and Sc during sediment migration  
379 and the preservation in marine depositional environments of their concentration distribution  
380 acquired in continental systems allow tracking source areas. In this study, we used the Hf–La/Th,  
381 La–Sc–Co/Th, and  $\Sigma$ REE–La/Yb indexes to identify the mineral composition of the source rock  
382 and their corresponding tectonic setting (**FIG. 11**). As shown by the Hf content ranging from 5.05  
383  $\times 10^{-6}$  to  $8.33 \times 10^{-6}$  and the La/Th ratio from 3.2 to 5.0., the lithology of the source rocks of the  
384 Bashkirian to Sakmarian deposits is mainly felsic (**FIG. 11a**). This is further attested by the Co/Th  
385 and La/Sc ratio of 0.47–1.93 and 2.73 – 8.40 respectively (**FIG. 11b**). The  $\Sigma$ REE ranges from  
386  $187.17 \times 10^{-6}$  to  $751.75 \times 10^{-6}$  and the La/Yb ratio from 1.06 to 2.80 (**FIG 11c**). Finally, the La–Th–  
387 Sc, Th–Co–Zr/10 and Th–Sc–Zr/10 plots in FIGS. 11d–e suggest that the tectonic setting of the  
388 source area was intermediate between an active continental margin and island arc environment.

389 (2) Vertical evolution of the REE and trace element patterns in relation with sea-level  
390 fluctuation

391 The  $\Sigma$ REE content of all analyzed samples ranges in  $(187.17\text{--}751.75) \times 10^{-6}$  and its average  
392 value is  $299.83 \times 10^{-6}$ , which is slightly higher than that in North American Shale Composite  
393 (NASC ( $173.1 \times 10^{-6}$ )).  $\Sigma$ LREE ranges from  $174.34 \times 10^{-6}$  to  $669.06 \times 10^{-6}$ , and  $\Sigma$ HREE from  
394  $(12.84 \times 10^{-6}$  to  $82.69 \times 10^{-6}$ ).  $\Sigma$ LREE/ $\Sigma$ HREE is 8.09–16.44, and the average value is 11.44, which  
395 is higher than that of NASC (7.50). The value of  $(\text{La}/\text{Sm})_N$  has a positive correlation with the  
396 fractionating degree of the LREE, whereas  $(\text{Gd}/\text{Yb})_N$  can reflect that of HREE. As *McLennan et al.*

397 (1990) pointed out, an enrichment in LREE indicates a subaqueous setting; otherwise, HREE  
398 showed an aquatic setting. From the Bashkirian to the Gzhelian, we interpret the increase in  
399  $\Sigma\text{LREE}/\Sigma\text{HREE}$  as indicative of a transgression phase. Similarly, the decrease in  $\Sigma\text{LREE}/\Sigma\text{HREE}$   
400 during the Asselian to the Sakmarian is interpreted as marine regression.

401 As mentioned above, the trace elements ratio can also serve as an index of sea level  
402 variations (Qiu *et al.*, 2015). During the early Pennsylvanian, the  $w(\text{Sr})/w(\text{Ba})$  of samples G3,  
403 DJ43, and DJ53 is over 0.5, and the  $w(\text{Ge})/w(\text{La})$  and  $w(\text{Rb})/w(\text{Zr})$  are over 1.5 and 0.75,  
404 respectively, which indicated a marine setting in the responding area. The  $w(\text{Ge})/w(\text{La})$  and  
405  $w(\text{Rb})/w(\text{Zr})$  of samples H4 and Y787 in the western EMA is less than 1.5 and 0.75, which  
406 represent a continental setting. Thus, we speculate that the main transgression prograded  
407 westward.

408 During the Asselian stage, the  $w(\text{Sr})/w(\text{Ba})$ ,  $w(\text{Ce})/w(\text{La})$ , and  $w(\text{Rb})/w(\text{Zr})$  of Y787 showed a  
409 declining trend, whereas the  $w(\text{V}) / (w(\text{Ni}) + w(\text{V}))$  and  $w(\text{CaO})/w(\text{MgO} \times \text{Al}_2\text{O}_3)$  indices increased  
410 to over 0.25 and 0.75. These results indicate that the sea level rose sharply during the Asselian  
411 stage; this conclusion is the similar to that of Xin *et al.* (2018). During the Sakmarian,  $w(\text{Sr})/w(\text{Ba})$   
412 in G3 decreased sharply to 0.2, indicating a regression process in the southern EMA. Combining  
413 the trace element ratio and the study of Zhu *et al.* (2008), we infer that the main regression was  
414 ESE directed.

#### 415 **6.4 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data**

416 The  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  data obtained on whole-rock samples show the fluctuation trend of the sea  
417 level caused by climate or the sediment rate change (**FIG. 14**). During the Pennsylvanian,  $\delta^{13}\text{C}_{\text{V-PDB}}$   
418  $\text{PDB}$  ranged from  $-8\text{‰}$  to  $-25\text{‰}$ , and  $\delta^{18}\text{O}_{\text{V-PDB}}$  ranged from  $-5\text{‰}$  to  $-15\text{‰}$ . The value shows a

419 negative excursion and a decreasing trend, indicating an warming climate and more rainfall. This  
420 result is consistent with the study of *Montanez et al. (2007)* and *Haq and Schutter (2008)*. During  
421 the Asselian–Sakmarian,  $\delta^{13}\text{C}_{\text{V-PDB}}\text{‰}(-4.35\text{‰}—1.36\text{‰})$  and  $\delta^{18}\text{O}_{\text{V-PDB}}\text{‰}(-9.42—11.35)$  showed  
422 a positive excursion, which speculate the sea level rising during this interval.

### 423 **6.5 Paleontological assemblage**

424 During the Pennsylvanian, the value of  $w_{\text{leaf}}$  (the amount of preserved leaf in macro-plant  
425 remnants) showed a rising trend from 0.47 to 0.52 including species such as *Cathaysiodendron*  
426 *nanpiaoense*, *Conchophyllum richthofenii*, *Alloiopteris sp.*, and *Lepidodendron cf. szeianumis*. In the  
427 early Pennsylvanian series, leaf remnants are generally fragmented and mixed with stems  
428 fragments (**FIG. 15**, samples G10 and DJ28). In the late Pennsylvanian, the  $w_{\text{leaf}}$  increased to  
429 0.52, suggesting a rising sea level; whereas the large proportion of complete leaf remnants (**FIG.**  
430 **15** G7 and H5) suggests a weakening hydrodynamic setting.

431 During the early Permian, the amount of plant fossils sharply increased, and the  $w_{\text{biomass}}$  of  
432 ferns such as *Lycopsids*, *Emplectopteris*, and *Emplectopteridium*, showed a sudden decrease during  
433 the transition from Kasimovian to Asselian. The root and stem fossils were predominant during  
434 the Asselian interval (**FIG. 15** G1), and the ratio of the broad leaf remnants( $w_{\text{leaf}}$ ) decreased to  
435 0.30. Up to the Sakmarian interval, the stem fragments keep predominant, whereas the  $w_{\text{leaf}}$   
436 declined to 22%.

### 437 **6.6 Clay minerals distribution**

438 As indicated above, continental sediment generally show a high chlorite and/or kaolinite  
439 content while marine sediment may show a high illite content. Furthermore, *Raucsik and Varga,*  
440 *(2008)* and *Chamley et al., (1993)* showed that differential settling of clay minerals during

441 sedimentation processes could explain the illite and smectite enrichment in marine settings away  
442 from the major river deltas and estuaries Triangle diagrams showing the proportion of each type  
443 of clay (chlorite and kaolinite are plotted on the same location) are thus use to infer the  
444 continental, shallow-marine or deep-marine depositional environment through time (FIG.16).

445 Northeastern part of the EMA: Wells DJ16 and DJ43 (**FIG. 16**) are characterized by a high  
446 chlorite content (>50%) in the Pennsylvanian, which shows an increasing in illite content  
447 characterizing the late Pennsylvanian (Kasimovian), which rise from 20% to 45%. whereas,  
448 chlorite and kaolinite are still predominant ( $w_{clay}$  (the ratio of  $w(\text{chlorite} + \text{kaolinite})/w(\text{illite})$ ) values  
449 are still over 1.0), indicate that the sediments from a continental setting is dominated in this  
450 interval. During the Asselian stage, the illite content of wells DJ16 and DJ43 decreased to 35%,  
451 further supporting the occurrence of a slight regression phase and an increasing input of  
452 continent-derived detrital material.

453 Southeastern part of the EMA: In wells DJ53 and DJ30 illite dominated (their content high up  
454 to 50%) during the Pennsylvanian to Sakmarian. All the above data indicate that the sediment  
455 deposited in the southern EMA may be below the sea level.

456 Southwestern part of the EMA: The Pennsylvanian section in well G3 shows a high chlorite  
457 content (>50%), and a  $w_{clay}$  index over 1.5, indicating that the southwestern part of the EMA is  
458 below the sea level up to the late Pennsylvanian. With the illite content decreasing dramatically  
459 (from 35% to 25%), we infer that a regression progress occurred in the southwestern block during  
460 the early Permian.

461 Northwestern part of the EMA: In well H4, chlorite and kaolinite dominate the clay spectrum  
462 from the Pennsylvanian to the Sakmarian, indicating a terrestrial depositional setting. As to the

463 Asselian–Sakmarian, the chlorite content in Y787 shows a decreasing trend (from 65% to 50%),  
464 indicating less sediment input from continental source regions.

## 465 **7. Provenance area identification and sediment supply calculation**

### 466 **7.1 Provenance**

#### 467 7.1.1 Provenance area during the Pennsylvanian

468 Based on published U-Pb data from *Yang et al. (2005)*, *Darby and Gehrels (2006)*, *Li et al.*  
469 *(2013)*, and *Sun et al. (2014)* we obtained the main characteristics of potential provenance areas  
470 in the Qinling orogenic belt (south), the Qaidam–Qilian orogenic belt (west) and the Yinshan  
471 orogenic belt (north)(**FIG. 8**). Furthermore, *Xie and Heller (2013)* pointed out that late Paleozoic  
472 uplift in the north China block may also be a potential source area for the Ordos basin. The age  
473 spectrum of JC-A and JC-B (early Pennsylvanian) show a major zircon U-Pb age population at  
474 762–974 Ma, similar to the one observed in the Qaidam–Qilian orogenic belt (**FIG. 8**). As for the  
475 peak point at 1750 Ma, we speculate that some sediment came from the Yinshan orogenic belt.  
476 According to **FIG. 11 d** and **f**, the trace elements pattern of the early Pennsylvanian samples  
477 suggest a source mainly composed of continental island arc and active continental margin  
478 lithologies, again suggesting a source in the Qaidam–Qilian and Yinshan orogenic belts (*Darby*  
479 *and Gehrels,2006*); *Golonka and Ford, 2000*). Finally, both HMA (samples DJ43 and DJ53) and  
480 ZTR indexes indicate a distal deposition process compatible with a source in the Qilian Shan and  
481 Yinshan (**FIG. 10**).

#### 482 7.1.2 Provenance area during the early Permian

483 The zircon U-Pb age population of 310–375 Ma recorded in the age spectrum of samples H5,  
484 DJ53, DJ51, and G2, is largely comparable to the one shown in samples from the Qinling orogenic

485 belts (*Sun et al., 2014; Xin et al., 2018*). Furthermore, *Zhu et al., 2011 and Dong et al., 2016*,  
486 indicated a source from the Qinling orogenic belt in the early Permian southern EMA deposits.  
487 Trace elements and REE patterns further indicate that the material was largely derived from  
488 Paleozoic sedimentary rock or igneous rock (as Sample E shown in **FIG. 12**) (*Shi et al., 2018*).  
489 Thus, we speculate that some of the early Permian sediment originated from the Qinling region.  
490 The Asselian record of wells H5 and DJ53 show two other zircon U-Pb age peaks at 1650–2300  
491 and 2300–2500 Ma. Based on the analogy with the age spectrum obtained by *Darby and*  
492 *Gehrels(2006)*we infer that some of the material also derived from the Yinshan orogenic belt. The  
493 increase observed in the ZTR index support that hypothesis of a more distal source.

494 In the Sakmarian stage, these two zircon U-Pb age peaks are absent from the age spectrum  
495 (**FIG. 8**), especially for the wells in the southern EMA (G2). Based on the HMA index (**FIG. 10**),  
496 we speculate that the provenance area of the sediment in the northern EDM is mainly the  
497 northern Ordos margin through long-distance transport. As to the southern EMA, the decreasing  
498 ZTR index indicates a switch from distal provenance to a more proximal source. Moreover, the  
499 detrital zircon U-Pb age spectrum and trace elements distribution (FIG 8 and 11) reflects material  
500 issued from the Yinshan and Qinling orogenic belt.

## 501 **7.2 Sediment supply rate calculation**

502 We selected two profiles to calculate the sediment deposition rate in the EMA. **FIGS. 17 and**  
503 **18** show the EMA in the N–S and E–S direction, respectively. We divided the profile into four  
504 layers: Penzi Formation (Ct–Cb), Taiyuan Formation (Cb–P2s), Upper Shanxi Formation (P2s–  
505 Pt10), and Lower Shanxi Formation (Pt10–Ps1).

### 506 **7.2.1 Pennsylvanian sedimentation rates**

507 As shown in the well calibration (**FIG. 1**), the Moscovian–Kasimovian stage in the early  
508 Pennsylvanian is represented by the Penzi Formation, while the Taiyuan Formation formed  
509 during the Gzhelian stage.

510 Then, we evaluated the strata thickness (A) on the basis of the seismic profile in **FIGS. 17** and  
511 **18**. During the Moscovian–Kasimovian stage (315.2–303Ma), the sediment from the eastern EMA  
512 was deposited, and its transport path showed a east direction as that in **FIG. 18**. The statistical  
513 mean thickness of the strata (A) in a series of the E–W direction profiles range from 70 m to 90 m.  
514 Thus, the average sedimentation rate at this stage is 8 m/Ma, whereas the minimum  
515 sedimentation rate is 5 m/Ma. During the Gzhelian stage, the decrease in the strata thickness to  
516 10–20 m corresponds to a declining trend of the sedimentation rate (2–4m/Ma) interval. The low  
517 sedimentation rate in the Gzhelian (*Jiang and Wang, 2012; Xu et al., 2015*) is directly related to  
518 the formation of the carbonate platform in the EMA, as the clastic input largely decreased.

#### 519 7.2.2 Early Permian sedimentation rates

520 The Asselian stage is represented by the Lower Shanxi Formation, and the Sakmarian by the  
521 Upper Shanxi Formation.

522 We selected a seismic profile in the 2D block, consistent with the sediment migration direction.  
523 According to the above study, sediment was mainly transported from the northern EMA during the  
524 Asselian stage (**FIG. 17**). During the Sakmarian, sediment was transported from two directions:  
525 the southern EMA was filled by proximal material from QOB(Qiling Orogenic belt), whereas the  
526 northern block was filled sediment from a distal area(YOB: Yinshan Orogenic belt). Thus, as the  
527 static data in **Table 3**, we evaluate the sedimentation rate in the southern EMA and calculated the  
528 sedimentation rate in the northern block on the basis of **Table.3**.

529 As shown in **FIG. 18**, the average strata thickness (A) of the Lower Shanxi Formation  
530 (Asselian298~295Ma) is 80–100 m, and the calculated sedimentation rate is 20–25 m/Ma. In the  
531 Upper Shanxi Formation (Sakmarian), more sediment were provided and the sedimentation rate  
532 reached 30 m/Ma in the southern block and 20~25 m/Ma in the north E.M.D. .

### 533 **7.3. Transgression/regression rate calculation**

#### 534 7.3.1 Transgression/regression rate during the Pennsylvanian

535 As shown in **FIGS.16a** and **16b**, the main transgression/regression direction during the  
536 Pennsylvanian is E–W in the northern block, whereas that in the southern block is N–NW.  
537 According to the profiles with a similar direction of transgression/regression as that of the  
538 northern block, the seismic packages (TP9–TP10) in **FIG. 18** are characterized by high continuity  
539 and low onlap . These characteristics indicate limited erosion during the sea level rising process.  
540 *Jiang and Wang. (2012) and Xu et al. (2015)* suggested that the transgression process in the  
541 Pennsylvanian was mainly controlled by accommodation changes. The average transgression  
542 rate during the Pennsylvanian was 15 km/Ma (T.P.1~4 and 8~10 **TABLE 3**). As to the southern  
543 block, the seismic packages showed an onlap from TP1 to TP3, indicating that the regional  
544 subsidence provided enough accommodation space for the sediment. Considering the distance  
545 from TP1 to TP3, the transgression rate on the southern block during the Pennsylvanian period  
546 was 7.6 km/Ma. Trace elements,  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$ , and paleontological assemblages show a sharp  
547 regression event during the late Pennsylvanian (Kasimovian) (**FIGS. 13, 14, and 15**). However,  
548 the resolution of the seismic profile makes it impossible to calculate the regression rate during  
549 that period.

#### 550 7.3.2 Transgression rate from the Asselian to the Sakmarian stage

551 **FIGURE 16c** shows an NW–W directed transgression during the Asselian stage. We selected  
552 **FIG. 18** to calculate the transgression rate. The TP13 and TP14 markers were chartered by a  
553 series of discontinuous seismic packages and distinct concave reflection boundaries. We  
554 calculated the average transgression of the EMA as 30 km/Ma. During the Sakmarian stage (**FIG.**  
555 **16d**), sea level variations occurred in a spatially variable pattern caused by the uplifting process  
556 in the southern orogenic belt. More proximal sedimentation lead to a sharp regression in the  
557 southern EMA. According to the seismic packages RP4, RP5, and RP6 (**FIG .17**), we obtained a  
558 regression rate of 50 km/Ma. In the northern block, the calculated average regression rate ranges  
559 from 15 km/Ma to 20 km/Ma according to the seismic packages RP9 and RP10 (**TABLE 3**).

## 560 **8. Mechanism of transgression or regression in intracratonic basin**

561 After calculating the sedimentation rates and the transgression/regression rates, we  
562 reconstructed the S2S system during the Carboniferous–Permian (**FIG. 19**) in accordance with  
563 the BQART model of *Syvitski and Milliman (2007)*. (**FIG. 20**) and classified it into three types: low-  
564 transgression rate and less-sediment supply S2S pattern (LT–LS) during the Pennsylvanian  
565 period, high-transgression rate and massive sediment supply S2S pattern (HT–MS), and low-  
566 transgression rate and massive sediment supply S2S pattern (LT–MS) during the Asselian to  
567 Sakmarian stage. Below, we infer the mechanism of Carboniferous–Permian transgression in  
568 accordance with the quantitative parameters of the S2S system.

### 569 **8.1 S2S distribution pattern between the interglacial and glacial epochs**

570 During the early Pennsylvanian (Bashkirian–Kasimovian), the EMA mainly developed tidal  
571 deposits. We distinguished the N-S directed tidal barriers, which developed on the former  
572 basement (*Shen et al.,2017*), separating the EMA from the western part of the Ordos Basin (both

573 sub-basins were evolving independently). In addition, the provenance study shows that the  
574 sediment came from a distal area in the western margin (**FIG. 19c**). Combining the above  
575 depositional setting, meandering channels in the western block and tidal channels in the eastern  
576 EMA (**FIG. 19a**). With the decrease in sedimentation rate and the increasing transgression area,  
577 the carbonate platform developed during the late Pennsylvanian period (Gzhelian) along the  
578 western EMA; the lithofacies also support this conclusion. Thus, we infer that the shallow-marine  
579 areas in the early stage. At the same period, the western sub-basin is characterized by floodplain  
580 and channel deposits (*Zhang et al. 1997*) associated a low-transgression rate and less-sediment  
581 supply (LT–LS S2S pattern) (**FIG. 20a**). Then, refer to the study of (*Michael et al.,*  
582 *2013; Hampson et al., 2014*), this pattern of S2S system further shows mass sediment attested  
583 by fine-grained deposits infills.

584 During the Asselian stage, the depositional setting constrained from seismic data shows a  
585 series of erosion channels (**FIG. 4 and 5**). At the same time regression continues in the western  
586 Ordos Basin, leading to shelf and floodplain depositional environments. *Ritts et al., (2006)*, and  
587 *Zhang et al., (1997)* pointed out that the facies association include braided- and  
588 meandering/anastomosed- channels (**FIG. 19**). The sedimentation and transgression/regression  
589 rates calculated above reveal that the high-transgression rate and mass sediment supply (HT–  
590 MS) (**FIG. 20b**) pattern was predominant during this interval. During the Sakmarian stage, the  
591 sediment input from the Yinshan orogenic belt declined slightly providing less coarse-grained  
592 material. We distinguished facies associations LFA.1 and LFA.3 during this interval and interpret  
593 it as a meandering channel. The S2S system that developed in the northern EMA had an HT–MS  
594 pattern. By contrast, the southern block developed various S2S systems. From the facies

595 association (FIG. 4), we conclude that delta fan or coarse-grained delta depositional  
 596 environments dominated during this interval. Thus, we infer that the Asselian–Sakmarian stage  
 597 with massive sediment input corresponds to the interglacial period. The S2S system in the  
 598 southern block shows a typical low-transgression rate and massive sediment supply (LT–MS)  
 599 S2S pattern (FIG. 21c). In the southern block, coarse-grained submarine fans or delta fans  
 600 developed in proximal areas. In the northern EMA, sedimentation occurred with more distal than  
 601 the southern one..

## 602 8.2 Mechanism of transgression from the perspective of the S2S system

603 Based on the depositional environments (FIG. 19) and reconstruction of the topography of the  
 604 basin derived from seismic data (FIG. 17 and 18), we propose the following S2S pattern model.

### 605 1 ) LT–LS S2S pattern (Pennsylvanian)

606 The topographic reconstruction of the Pennsylvanian basin indicates low continental slopes.  
 607 The stratigraphy indicates a limited carbonate platform and shallow waters characterized by tidal-  
 608 influenced deposits. Then, we calculated the parameters (TABLE 3) of the LT–LS S2S pattern  
 609 during the Pennsylvanian and built the response equations as follows:

$$610 \quad V = aA + bS + c\alpha \cdot e^H + m, \quad (\text{FIG. 20d})$$

611 where  $V$  is the volume of the S2S system, which can be obtained from the seismic data;  $A$  is  
 612 the distance between the provenance area and the sink area;  $S$  represents the sedimentation  
 613 rate;  $\alpha$  shows the dip angle of the slope; and  $H$  is the relative sea level.  $a$ ,  $b$ , and  $c$  in the  
 614 equation are obtained in accordance with the fitting of the data in TABLE 3.

615 As stated above, the volume of the S2S system during the Pennsylvanian is limited, and most  
 616 sediment accumulated on the slope (FIG. 20a). The bubble chart shows that  $\alpha$  is inversely

617 proportional to the relative sea level (H). We infer that the key factor controlling the relative sea (H)  
 618 level is the variations in slope dip ( $\alpha$ ). *Lan et al. (2016)* also pointed out that the spatial variability  
 619 transgression direction is caused by topographic variations. On this basis, as the general slope  
 620 (shown as the morphology map in **FIG. 16**), we can explain the fast transgression in the northern  
 621 EMA and the slow transgression rate in the southern block.

## 622 2 ) HT/LT–MS S2S pattern

623 The facies association pattern in FIG. 19 shows a spatially variable evolution in the south and  
 624 north block of the EMA during the Asselian–Sakmarian. As to the northern part, the stratigraphy  
 625 indicates braided channels or meandering/anastomosed rivers. The southern part was dominated  
 626 by fan deltas or alluvial fans. Then, we statistically calculated the S2S parameters in the  
 627 Asselian–Sakmarian (**TABLE 3**) and obtained the following fitting equation:

$$628 \quad V = a\alpha \cdot A \cdot H + be^S + m. \quad (\text{Fig. 20e})$$

629 In the HT–MS system, the volume of the S2S system is affected by the sedimentary supply (S)  
 630 and the relative sea level (H). For the Sakmarian interval, we assumed that V is a determined  
 631 value. The relative sea level (H) is inversely proportional to the sedimentary supply (S).  
 632 Furthermore, *Zhu et al., (2008)* pointed the subsidence rate keep stable during this interval. Thus,  
 633 we speculate that in the HT–MS system the sedimentation rate is the key factor driving the sea  
 634 level.

635 As to the LT–MS system in the southern block, we statistically calculated the S2S parameters  
 636 during the Sakmarian and obtained

$$637 \quad V = a\alpha \cdot A \cdot H \cdot S + m. \quad (\text{Fig. 20f})$$

638 The bubble chart in **FIG. 20 f** shows that the accumulated sediment and the dip angle can

639 affect sea-level fluctuation as generally inferred in sequence stratigraphy studies (*Zhu et al.*,  
640 2008). Due to the uplift caused by the formation of the Qinling orogenic belt, the slow  
641 transgression rate, and the fast regression rate during the Sakmarian were mainly controlled by  
642 the substantial amount of sediment accumulated and the increase in dip angle.

## 643 9. Conclusion

644 1. During the Pennsylvanian, sediments mostly came from the Qaidam–Qilian orogenic belt  
645 and were deposited in a distal setting. Some distal sediments from the northern margin of the  
646 Ordos Basin were injected in the dispersal area. The sediment accumulation rate during this  
647 interval ranges from 5 m/Ma to 10 m/Ma. During the Asselian–Sakmarian interval, material was  
648 derived from multiple provenance areas. The sediment from the Qinling orogenic belt  
649 accumulated at a rate of 30 m/Ma, and that from the western and northern Ordos Basin  
650 accumulated at 10 and 30 m/Ma, respectively. The timing of sedimentation rate changes is  
651 synchronous with the interglacial–glacial process in the early Permian.

652 2. During the early Pennsylvanian, transgression occurred towards the NEE, and the relative  
653 horizontal transgression rate reached as high as 15 km/Ma. We classified this kind of S2S pattern  
654 as LT–LS. During the Asselian stage, transgression occurred toward the NNW with a rate of 30  
655 km/Ma. During the Sakmarian, the sea level rose again, and the direction of transgression in the  
656 EMA was NW. The transgression rate was up to 50 km/Ma. We also identified two different S2S  
657 patterns, that is, LT–MS in the south and HT–MS in the north.

658 3. In the LT–LS pattern, the transgression/regression process was dominated by the dip angle  
659 of the morphology due to lower sediment supply during the glacial period (Pennsylvanian). In this  
660 pattern, the intensive reworking flow determined the region where coarse-grained sediments

661 developed. As the S2S pattern transitions to the HT–MS or the LT–MS pattern, we speculate that  
662 the migration of the sediment distance is a key factor that influences the transgression/regression  
663 process.

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672

## 673 **Figure Captions**

### 674 **FIG. 1**

675 Geomorphological map of the North China Block. a. Faults developed in the Ordos Basin. The  
676 area with the dotted line is the EMA in the Ordos Basin. The generalized stratigraphic column is  
677 shown in the left. Seismic data calibration and the stratigraphic marker referred to the study of *Zhu*  
678 *et al. (2008)*. From bottom to top: Penci Formation, Taiyuan Formation, Shanxi Formation, and  
679 Shihhotse Formation.

### 680 **FIG. 2**

681 The stratigraphy developed in Y787, DJ53, DJ43, H4, and G3. From left to right of each well is  
682 the Grammy curve, lithology. With the zircon age data shown in DJ53 and the related reference

683 (*Sun et al., 2014; Sun et al., 2013; Wang et al., 2003*), we identified the deposited age of layers,  
684 including the Penzi Fm. Taiyuan Fm, Lower Shanxi Fm. and Upper Shanxi Fm. Litho.=Lithology,  
685 Chronolo.=Geochronology.

686 **FIG. 3**

687 Lithofacies templates. Seven kinds of lithofacies associations are provided on the right side.  
688 From left to right: slow-rate one-directional traction flow deposited; high-rate one-directional  
689 tractional flow, slow-rate bidirectional traction flow, mass flow deposited, static water deposited  
690 and carbonated plate deposited.

691 **FIG. 4**

692 From left to right shows the seismic profile, well log of G6 and their responding cumulative  
693 grain-size curve. We interpreted the lithofacies according to the core sample calibration. The  
694 sandstone can be identified by the low GR value, while the limestone is characterized by the low  
695 DEN value. Furthermore, integrated the well log and the seismic reflection, we interpreted the  
696 facies association in the Pennsylvanian, including the tidal channel and the barriers.

697 **FIG. 5**

698 From left to right shows the seismic profile, well log of DJ12 and their responding cumulative  
699 grain-size curve. The cross-bedding in LFA.1 and LFA.2 indicated a tractional flow developed.  
700 Calibrated these on the seismic data, we recognized a series of the erosion channel developed.

701 **FIG. 6**

702 Representative zircon CL images. The small white circles represent the sites of U–Pb age  
703 analyses, and the measured ages are shown in Ma.

704 **FIG. 7**

705 Concordia plot of the detrital zircon of JC-A and JC-B (a), H5 and DJ53 (b), G2 and DJ51 (c).  
706 Data influenced by Pb loss and inclusion are deleted. Only analyses with concordances of >90%  
707 are included in the combined probability density plot and histogram and discussed in the  
708 subsequent sections.

709 **FIG. 8**

710 Zircon U–Pb age spectra of the Paleocene sediment in JC-A, JC-B, H5, DJ53, DJ51, and G2.  
711 Potential source area: Qinling orogenic belt, Qaidam–Qilian orogenic belt, Yinshan orogenic belt,  
712 West North China Block (*Xie and Heller, 2013; Kröner et al., 2006; Darby and Gehrels, 2006; Zhu*  
713 *et al., 2011*).

714 **FIG. 9**

715 As *Hoskin and Black (2000)* and *Teipel et al. (2004)* pointed out the according to the ratio of  
716 Th/U value, we can infer the sediment are mainly provided by the igneous sources (Th/U over  
717 0.5), or the metamorphic sources (Th/U less than 0.1). And the Th/U value between the 0.2 and  
718 0.5 indicate a mixture of the igneous and metamorphic source. Intersection diagrams for the Th/U  
719 ratio and age distribution. From left to right: samples of the Bashirian–Gzhelian, Asselian, and  
720 Samarian stages.

721 **FIG. 10**

722 HMA of Y521, H4, Y787, Y653, G3, Y432, DJ43, and DJ53. We classified the HMA into five  
723 types: rutile–prominent (rutile>60%), pyrite–prominent (pyrite>60%), titanite–prominent  
724 (titanite>50%), tourmaline–prominent (tourmaline>30%), and zircon–prominent (zircon>30%)  
725 EMA=East Margin area of Ordos Basin

726 **FIG. 11**

727 Points in the source rock discrimination chart show the tectonic setting of the EMA sediment.  
728 The first line shows Hf–La/Th (a), La/Sc–Co/Th (b), and  $\Sigma$ REE–La/Yb (c). The second line from  
729 left to right shows the La–Th–Sc, Th–Co–Zr, and Th–Sc–Zr.

730 **FIG. 12**

731 (a) Geological map of the Ordos Basin. We selected samples A, B, C, D, E, F, and G (b)  
732 around the EMA to study their clay mineral distribution. (c) represents the REE distribution in the  
733 Pennsylvanian with data from DJ28, DJ43, and DJ53.  $\Sigma$ LREE/ $\Sigma$ HREE is 8.09–16.44/11.44; (d)  
734 and (c) are the REE distribution in the Asselian and Sakmarian, respectively. We obtained the  
735 value of  $\Sigma$ LREE/ $\Sigma$ HREE in the Asselian is 9.74–15.72/10.12, whereas its value in the Sakmarian  
736 is 8.01–11.62/9.12. These values indicate the regression process during the Asselian and  
737 Sakmarian.

738 **FIG. 13**

739 Trace element vertical distribution in G3, H4, DJ43, DJ53, and Y787.

740 **FIG. 14**

741  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  data vertical distribution in Y787, DJ53, DJ 43, H4, and G3.

742 **FIG. 15**

743 Column from left to right show the  $w_{\text{leaf}}$  index,  $w_{\text{biomass}}$  index, micropaleontological assemblage  
744 (*Pfefferkorn and Wang, 2007*), and the plant fossils in the core samples. Referring to the  
745 statistical data of paleontological assemblage, we found that the sea level of the EMA increased  
746 during the Pennsylvanian and declined sharply during the Asselian–Sakmarian.

747 **FIG. 16**

748 We built the paleomorphology on the basis of 2D seismic data and the relative sea level as the

749  $w_{clay}$  index. From bottom to top: paleo-sea level developed from the early Pennsylvanian (a), the  
750 late Pennsylvanian (b), the Asselian (c), and the Sakmarian (d).

751 **FIG. 17**

752 The seismic profile in the N–S direction and the location is shown in **FIG. 1**. We distinguished  
753 a series of tidal channel with the swing channels as the concave of the seismic event. As to the  
754 late Pennsylvanian (Gzhelian), the limestone developed and showed a continuous and high  
755 amplitude reflection. Referring to *Zecchin et al. (2008)*, we distinguished a series of forest seismic  
756 packages and numbered them as TP1 to RP6.

757 **FIG. 18**

758 The seismic profile in the W–E direction and the location is shown in **FIG. 1**. We distinguished  
759 a series of forest seismic packages and numbered them as TP8 to RP10.

760 **FIG. 19**

761 (a) to (c) show the patterns of the three stages of the S2S system evolution during the  
762 Pennsylvanian and Asselian to Sakmarian. During the first stage, sediment migrated from a distal  
763 area, and a barrier developed on the slope. With the provenance area transfer, the north Yinshan  
764 orogenic provided mass sediment in this interval. During the third stage, the Qinling orogenic belt  
765 formed and lead delta fan developed in the southern block. We classified the S2S system of the  
766 first stage as a low-transgression-rate and less-sediment-supply S2S pattern. The second stage  
767 and the sediment rounding system in the northern block during the third stage has a high-  
768 transgression-rate and mass-sediment-supply S2S pattern, whereas the southern block in the  
769 third stage has a low-transgression-rate and mass-sediment-supply S2S pattern.

770 **FIG. 20**

771 S2S parameters (a to c) marked in the sections, including the bed evaluation height, the  
772 sediment supply and the relative sea level. Then, we fit these parameters (in **TABLE 3**) and  
773 obtained the balloon chart (d to f). The diameter represents the volume of the S2S system, and  
774 we can speculate the mechanism for the transgression/regression process in accordance with the  
775 fitting equation.

## 776 **Table Caption**

### 777 **Table 1**

778 Facies association descriptions.

### 779 **Table 2**

780 Heavy mineral assemblage in the potential source.

### 781 **Table 3**

782 Parameters of the S2S system. Statistical data were based on seismic data. We obtained the  
783 dip angle with the slope dip along the slope.

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TABLE 1

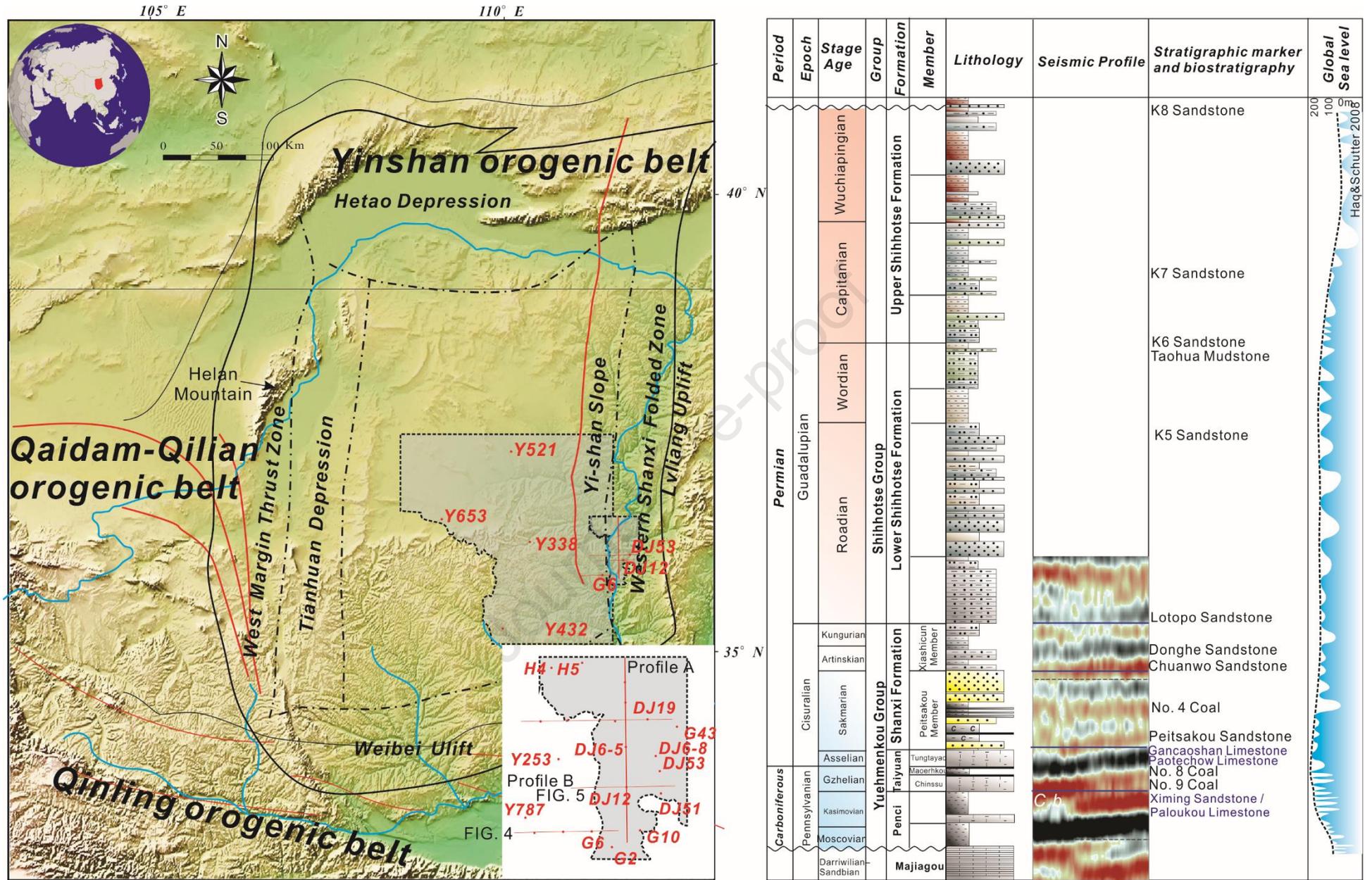
<b>Facies Association</b>	<b>Lithofacies</b>	<b>Main sedimentary features</b>	<b>Inferred depositional environment</b>
LFA.1	<i>Sp-Sm-M</i>	Moderately to well sorting thick sandstone. There are thick bad association trough crossing bedding. The thin layer of mudstone observed in the top.	Slow-rate flow deposited (Meandering channel environment) (refer to <i>Blum and Törnqvist (2000)</i> )
LFA.2	<i>Sb-Scm(Scc)-M</i>	Moderately sorted homogeneous sandstone. Regularly lamellar crossing bedding in the top.	High-rate traction flow deposited (Braid channel environment)
LFA.3	<i>Sb-Scm-Scc-M</i>	Strong rhythmic alteration of darker or lighter layers composed of fine-grained slit stone. The double-clay cross-bedding in the upper part.	Bidirectional slow-rate traction flow deposit(Tidal environment)
LFA.4	<i>Gp-Sd-Gm</i>	Gravel-supported conglomerate with regular imbedded layers. Upper to the top, matrix-supported conglomerate shows an imbricated orientation	A static water setting (Swamp)(refer to <i>Ichaso and Dalrymple (2009)</i> )
LFA.5	<i>Sd-Scc-M</i>	Moderately sorted fine stone in the bottom. Deformation structure occurs in the laminae. Despite intense bioturbation, there is no single trace identified.	Interchannel environment
LFA.6	-	Homogeneous carbonate with some mud clast identified in the top	Carbonate plate environment

TABLE.2

Period	Potential Source Area	Zircon	Tourmaline	Rutile	Titanite	Garnet	Ilmenite	Apatite	Pyroxene	Pyrite	Spinel	Types of heavy mineral assemblage	References
Pennsylvanian	Yishan Orogenic belt	28.25	28.96	9.28	12.00			10.08			3.36	HMA-I Rutile dominated	<a href="#">Chen et al., 2009</a>
Pennsylvanian	Qaidam–Qilian orogenic belt	9.08	0.04	0.15	0.01			7.96		58.6		HMA-II Pyrite dominated	<a href="#">Du et al., 2017</a>
Early Permian	Yishan Orogenic belt	20.91	1.93	11.25	30.00			7.96				HMA-III Titanite dominated	<a href="#">Ruppen et al., 2014</a> ; <a href="#">Li et al., 2014</a>
Early Permian	Yishan Orogenic belt	10.96	36.15	3.94	7.02			0.99				HMA-IV Tourmaline dominated	<a href="#">Xin et al., 2018</a>
Early Permian	Qiling orogenic belt	43.42	5.74	11.48	16.86		0.12		0.12		0.12	HMA-V Zircon dominated	<a href="#">Zhang et al., 2009</a>

TABLE 3

<i>Fm.</i>	<i>Block</i>	<i>Ave. Dip Angle</i>	<i>Provenance Area</i>	<i>Sink Area</i>	<i>Pathway Cut Area/km<sup>2</sup></i>	<i>Uplift Evaluated Height</i>	<i>Average Transgression/Regression Rate</i>
	<i>T.P.1</i>	<i>0.74</i>	<i>11.13</i>	<i>10.09</i>	<i>9.78</i>	<i>76.13</i>	<i>13.50</i>
	<i>T.P.2</i>	<i>0.94</i>	<i>11.54</i>	<i>12.86</i>	<i>7.36</i>	<i>73.23</i>	<i>14.09</i>
	<i>T.P.3</i>	<i>0.75</i>	<i>8.92</i>	<i>12.87</i>	<i>8.48</i>	<i>75.03</i>	<i>14.89</i>
<i>Penzi Formation</i>	<i>T.P.4</i>	<i>0.86</i>	<i>11.60</i>	<i>11.82</i>	<i>5.70</i>	<i>70.74</i>	<i>14.45</i>
	<i>R.P.1</i>	<i>0.80</i>	<i>8.93</i>	<i>16.68</i>	<i>6.80</i>	<i>74.57</i>	<i>15.80</i>
	<i>R.P.2</i>	<i>0.71</i>	<i>7.68</i>	<i>16.24</i>	<i>6.90</i>	<i>72.18</i>	<i>13.40</i>
	<i>R.P.3</i>	<i>0.87</i>	<i>9.98</i>	<i>9.27</i>	<i>7.04</i>	<i>72.76</i>	<i>16.36</i>
<i>Taiyuan Formation</i>	<i>T.P.4</i>	<i>0.81</i>	<i>11.94</i>	<i>12.14</i>	<i>8.72</i>	<i>53.36</i>	<i>17.96</i>
	<i>T.P.5</i>	<i>0.97</i>	<i>16.24</i>	<i>16.68</i>	<i>8.76</i>	<i>53.10</i>	<i>13.17</i>
<i>Lower Shanxi Formation</i>	<i>T.P.6</i>	<i>0.79</i>	<i>20.90</i>	<i>21.49</i>	<i>7.12</i>	<i>73.18</i>	<i>44.10</i>
	<i>R.P.4</i>	<i>0.68</i>	<i>26.53</i>	<i>26.34</i>	<i>7.34</i>	<i>72.54</i>	<i>53.06</i>
<i>Upper Shanxi Formation</i>	<i>R.P.5</i>	<i>0.83</i>	<i>11.24</i>	<i>10.21</i>	<i>8.46</i>	<i>80.35</i>	<i>25.04</i>
	<i>R.P.6</i>	<i>0.88</i>	<i>16.30</i>	<i>16.91</i>	<i>7.20</i>	<i>84.20</i>	<i>37.73</i>
	<i>T.P.7</i>	<i>0.97</i>	<i>22.05</i>	<i>20.28</i>	<i>6.78</i>	<i>79.57</i>	<i>32.20</i>
	<i>T.P.8</i>	<i>0.84</i>	<i>8.26</i>	<i>17.38</i>	<i>8.24</i>	<i>77.08</i>	<i>17.32</i>
	<i>T.P.9</i>	<i>0.94</i>	<i>8.53</i>	<i>8.25</i>	<i>6.84</i>	<i>70.83</i>	<i>16.55</i>
<i>Penzi Formation</i>	<i>T.P.10</i>	<i>0.89</i>	<i>8.67</i>	<i>9.50</i>	<i>10.44</i>	<i>78.00</i>	<i>16.72</i>
	<i>R.P.7</i>	<i>0.69</i>	<i>9.88</i>	<i>10.23</i>	<i>7.28</i>	<i>70.93</i>	<i>23.86</i>
	<i>R.P.8</i>	<i>0.94</i>	<i>8.75</i>	<i>8.59</i>	<i>7.04</i>	<i>75.81</i>	<i>15.12</i>
<i>Taiyuan Formation</i>	<i>T.P.11</i>	<i>0.87</i>	<i>15.76</i>	<i>16.18</i>	<i>9.02</i>	<i>53.17</i>	<i>19.88</i>
	<i>T.P.12</i>	<i>0.84</i>	<i>10.33</i>	<i>8.33</i>	<i>8.56</i>	<i>53.10</i>	<i>19.82</i>
<i>Lower Shanxi Formation</i>	<i>T.P.13</i>	<i>0.64</i>	<i>5.78</i>	<i>5.79</i>	<i>9.68</i>	<i>74.97</i>	<i>21.57</i>
	<i>R.P.9</i>	<i>0.67</i>	<i>5.42</i>	<i>6.46</i>	<i>7.30</i>	<i>72.19</i>	<i>27.41</i>
<i>Upper Shanxi Formation</i>	<i>R.P.10</i>	<i>0.58</i>	<i>6.16</i>	<i>5.14</i>	<i>6.34</i>	<i>84.91</i>	<i>37.71</i>



Period	Epoch	Stage Age	Group	Formation	Member	Lithology	Seismic Profile	Stratigraphic marker and biostratigraphy	Global Sea level
Permian	Guadalupian	Wuchiapingian	Shihhotse Group	Upper Shihhotse Formation				K8 Sandstone	200 100 0 -100 -200 Haq & Schutter 2008
		Capitanian					K7 Sandstone		
		Wordian					K6 Sandstone Taohua Mudstone		
	Roadian	Shihhotse Group	Lower Shihhotse Formation			K5 Sandstone			
						Lotopo Sandstone			
						Donghe Sandstone Chuanwo Sandstone			
Carboniferous	Cisuralian	Kungurian	Yuehmenkou Group	Shanxi Formation	Xiaoshicun Member			No. 4 Coal	
		Artinskian					Peitsakou Sandstone		
		Sakmarian			Peitsakou Member		Gancaoshan Limestone Paotechow Limestone No. 8 Coal No. 9 Coal		
	Pennsylvanian	Pencu	Asselian	Tunglayac Member				Ximing Sandstone / Paloukou Limestone	
			Gzhelian	Macerhoku					
			Kasimovian	Chinssu					
Darriwilian-Sandbian			Majiagou						



**FIG 2**

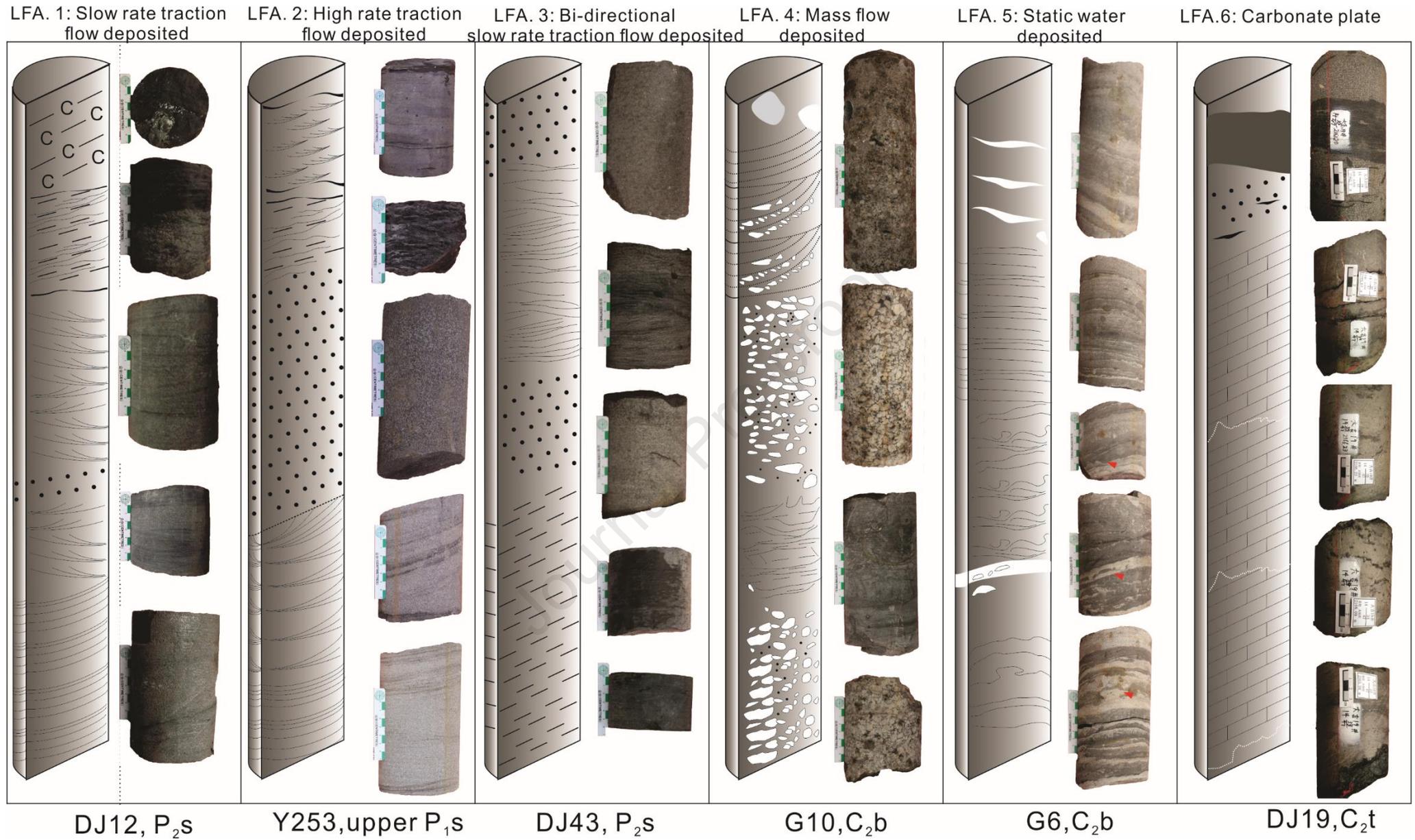


FIG 3

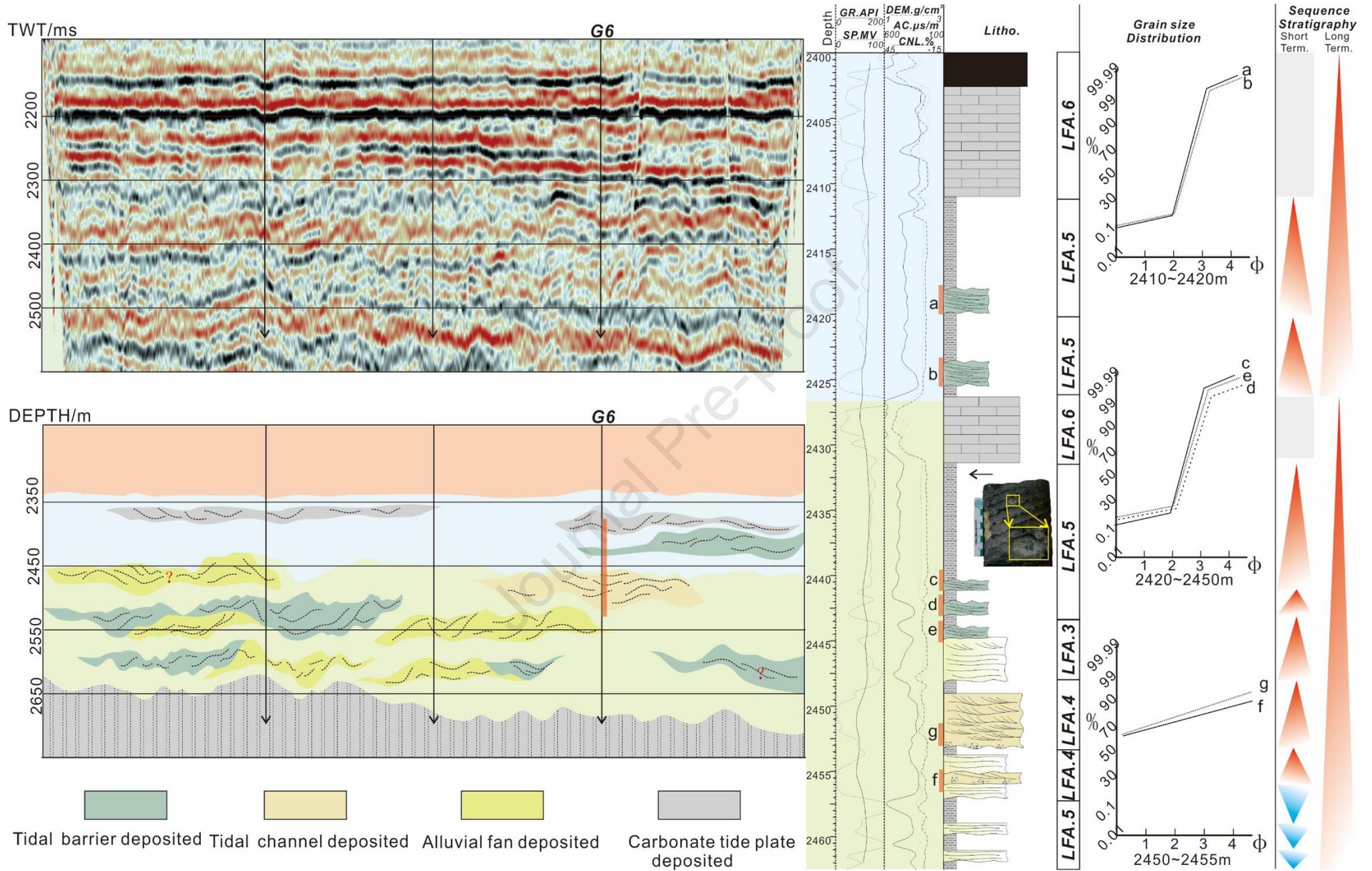


FIG 4

Tidal barrier deposited    Tidal channel deposited    Alluvial fan deposited    Carbonate tide plate deposited

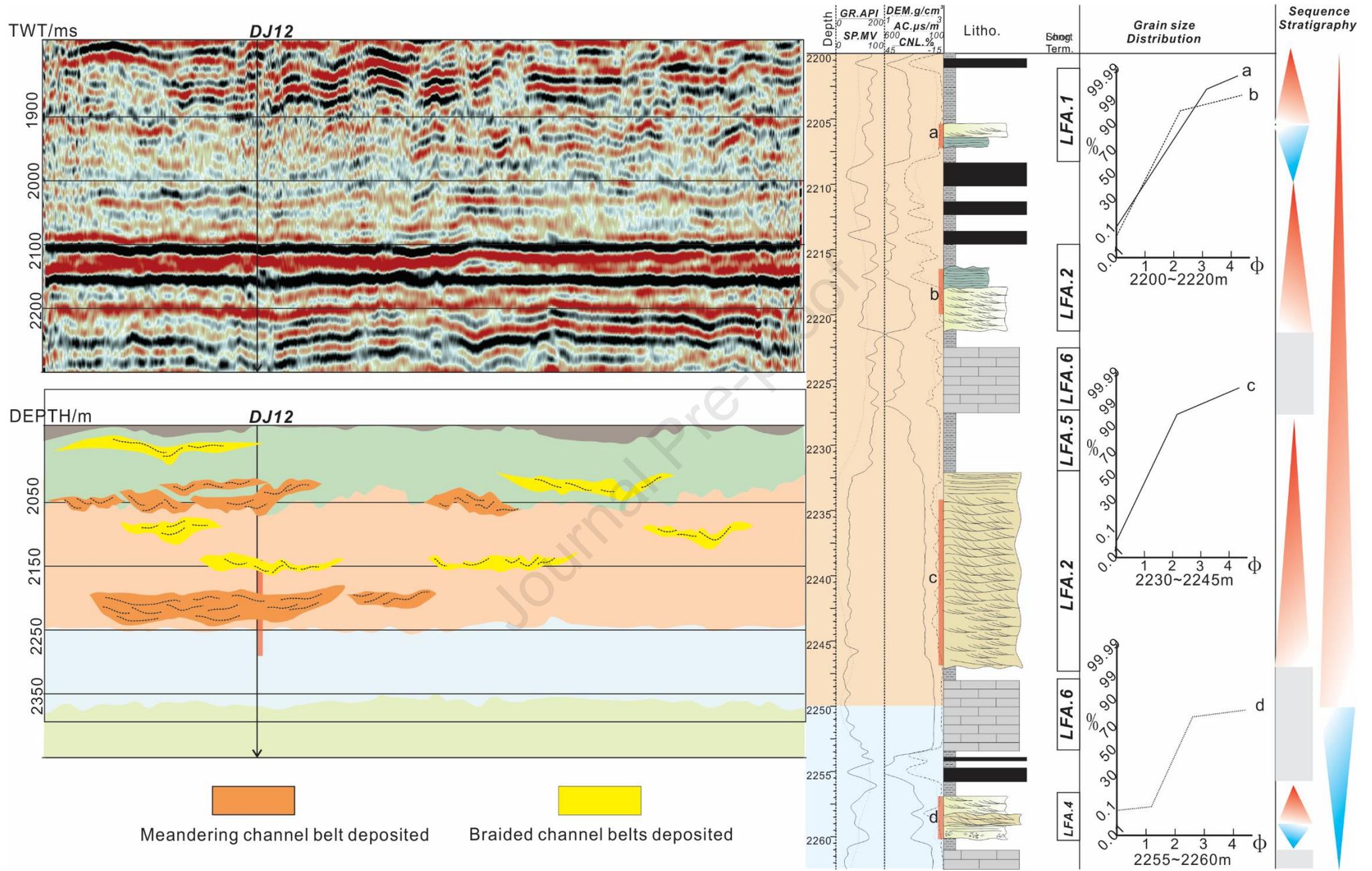
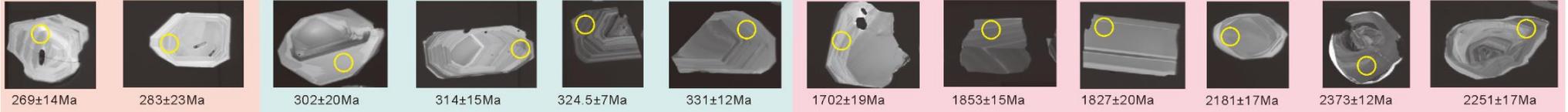


FIG 5

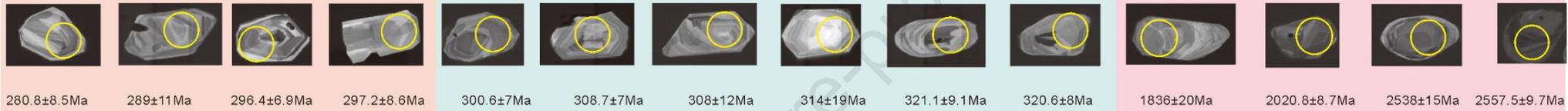
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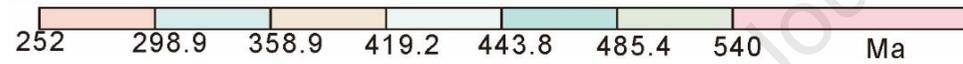


FIG 6

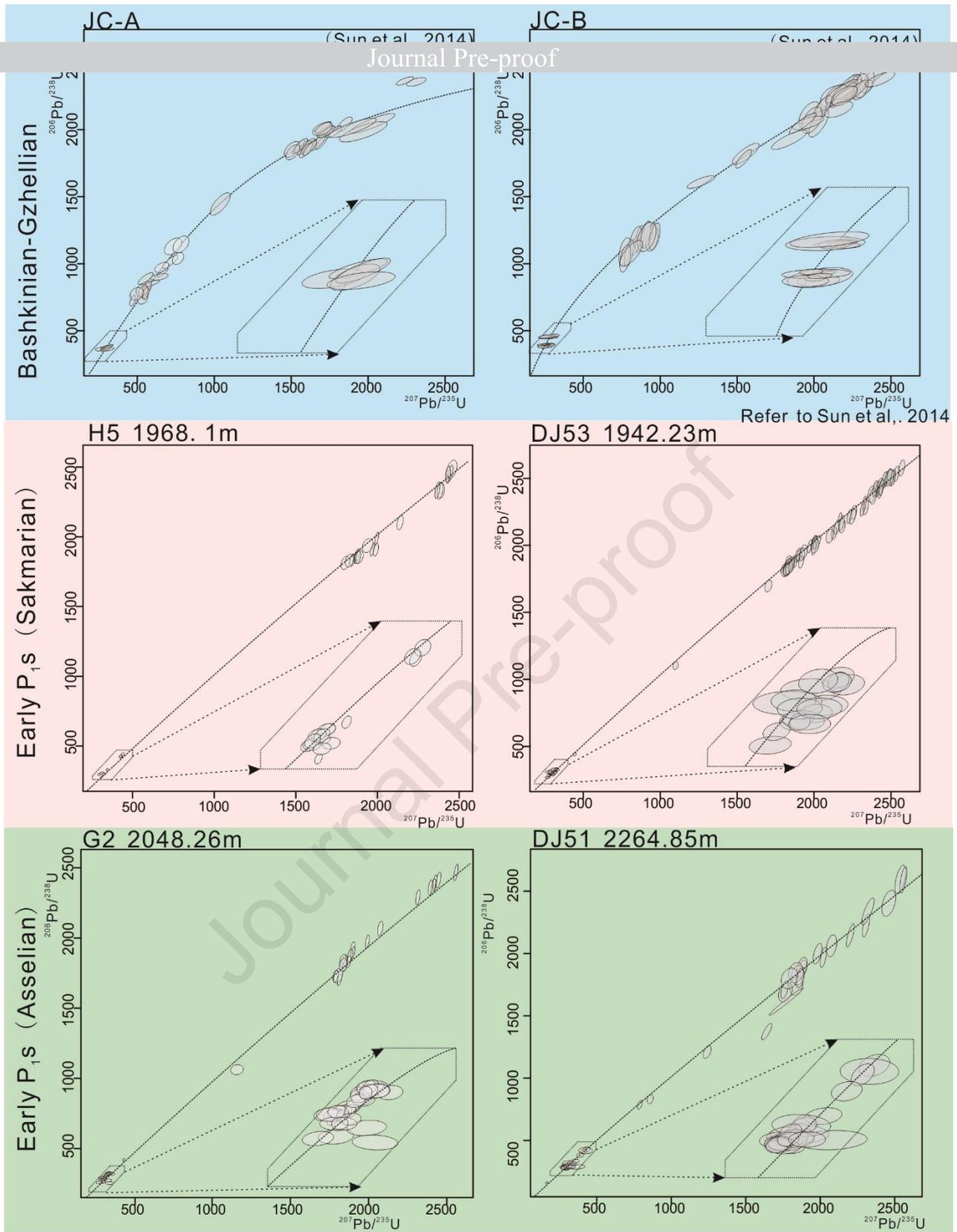


FIG 7

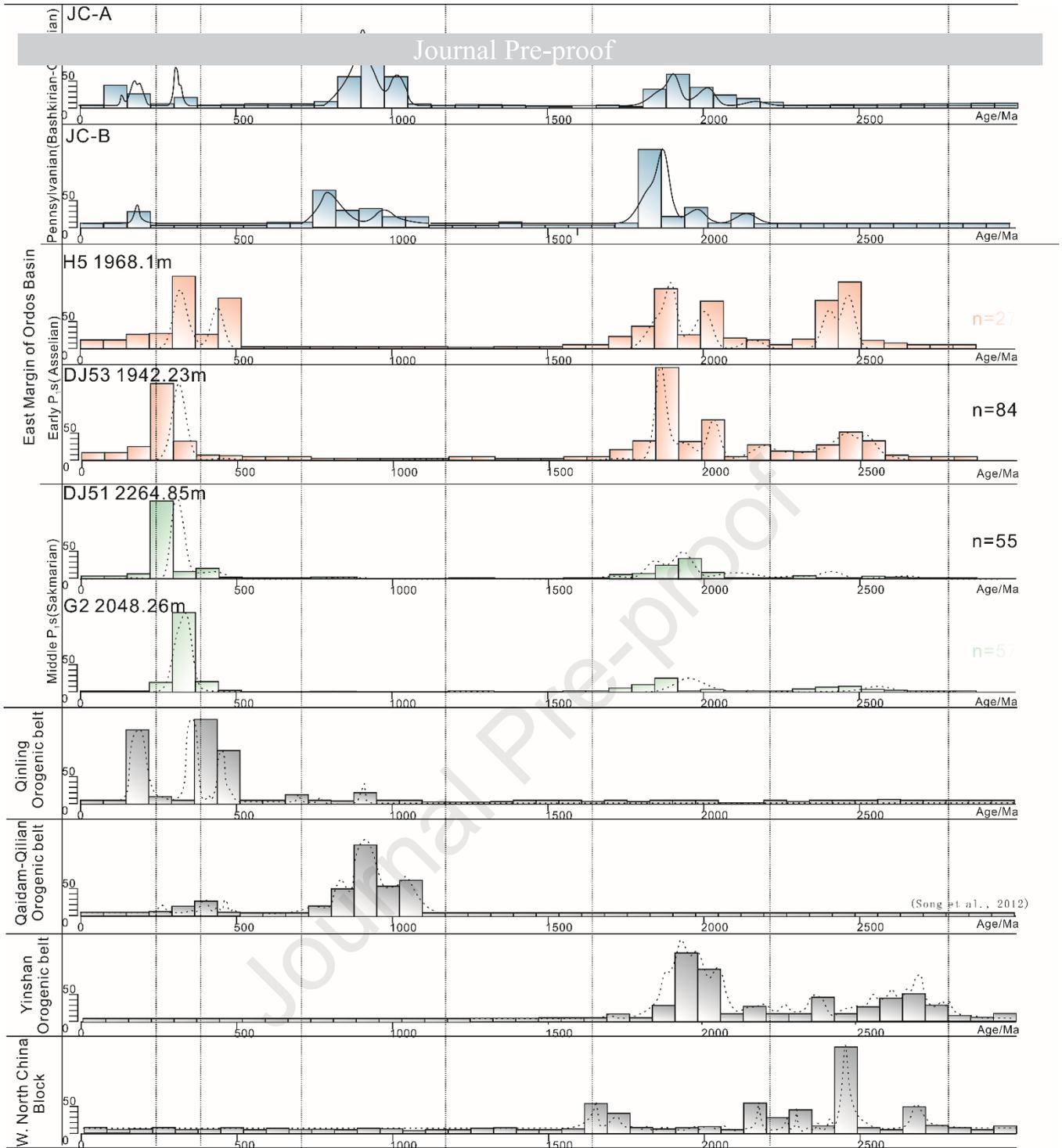


FIG 8

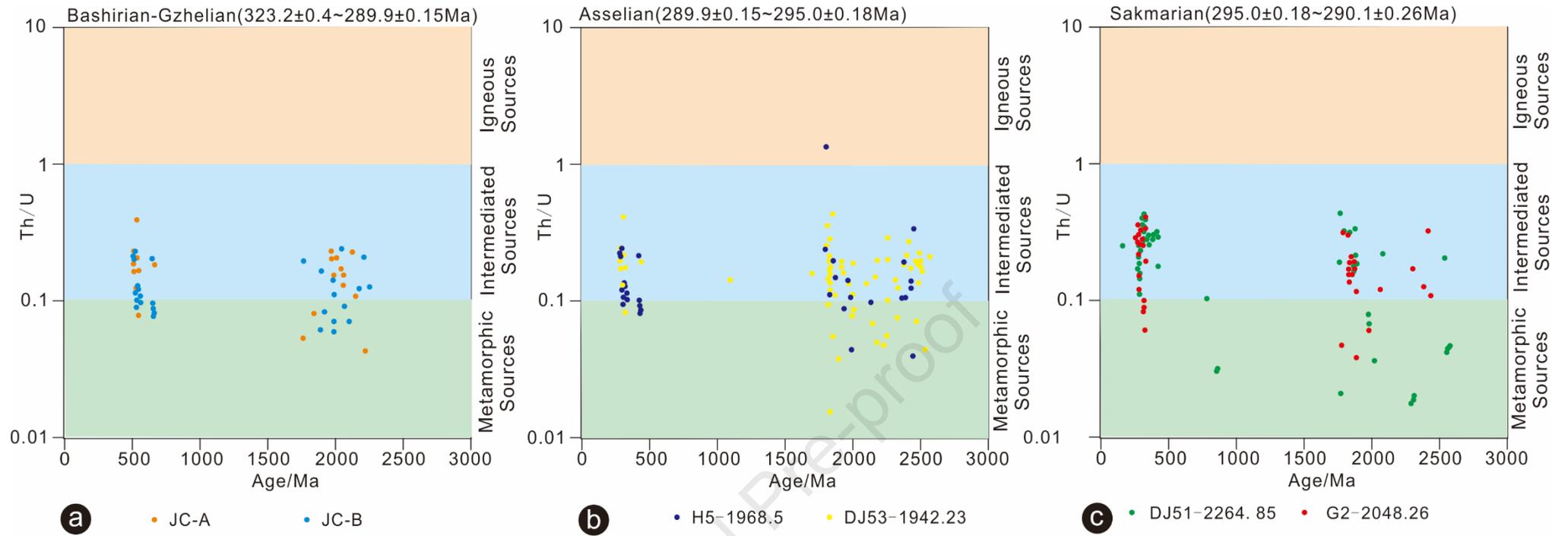


FIG 9

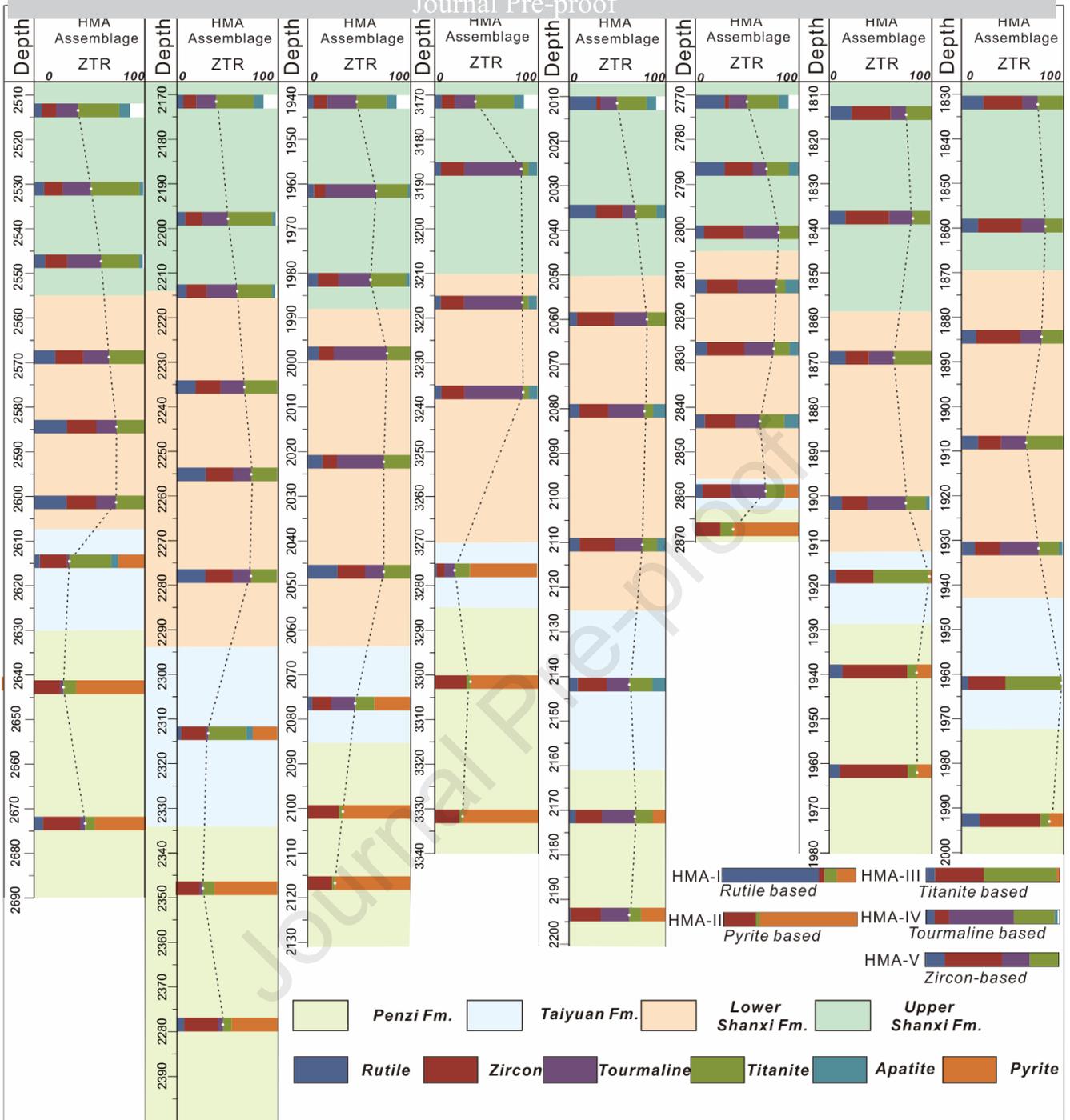


FIG 10

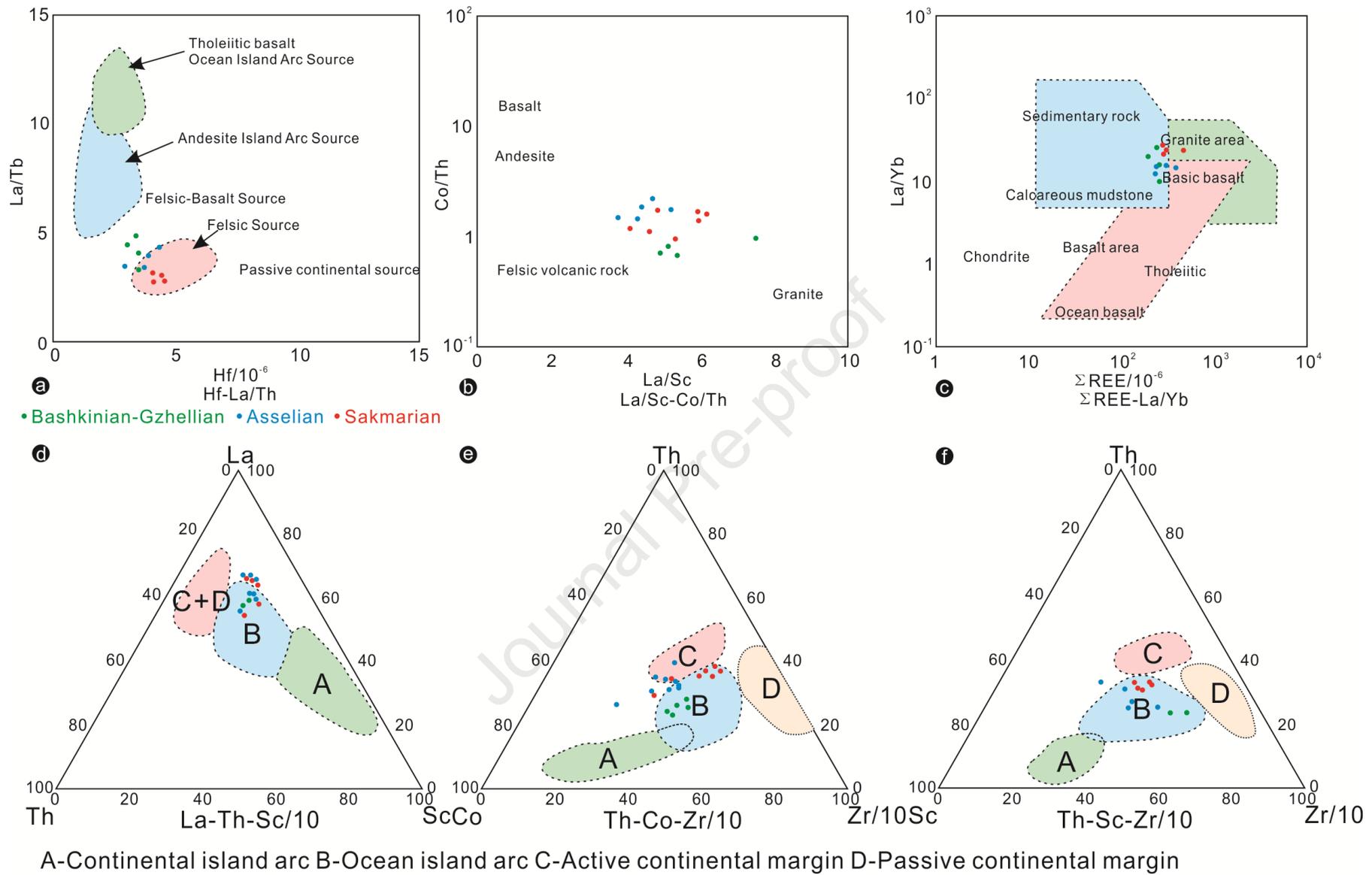


FIG 11

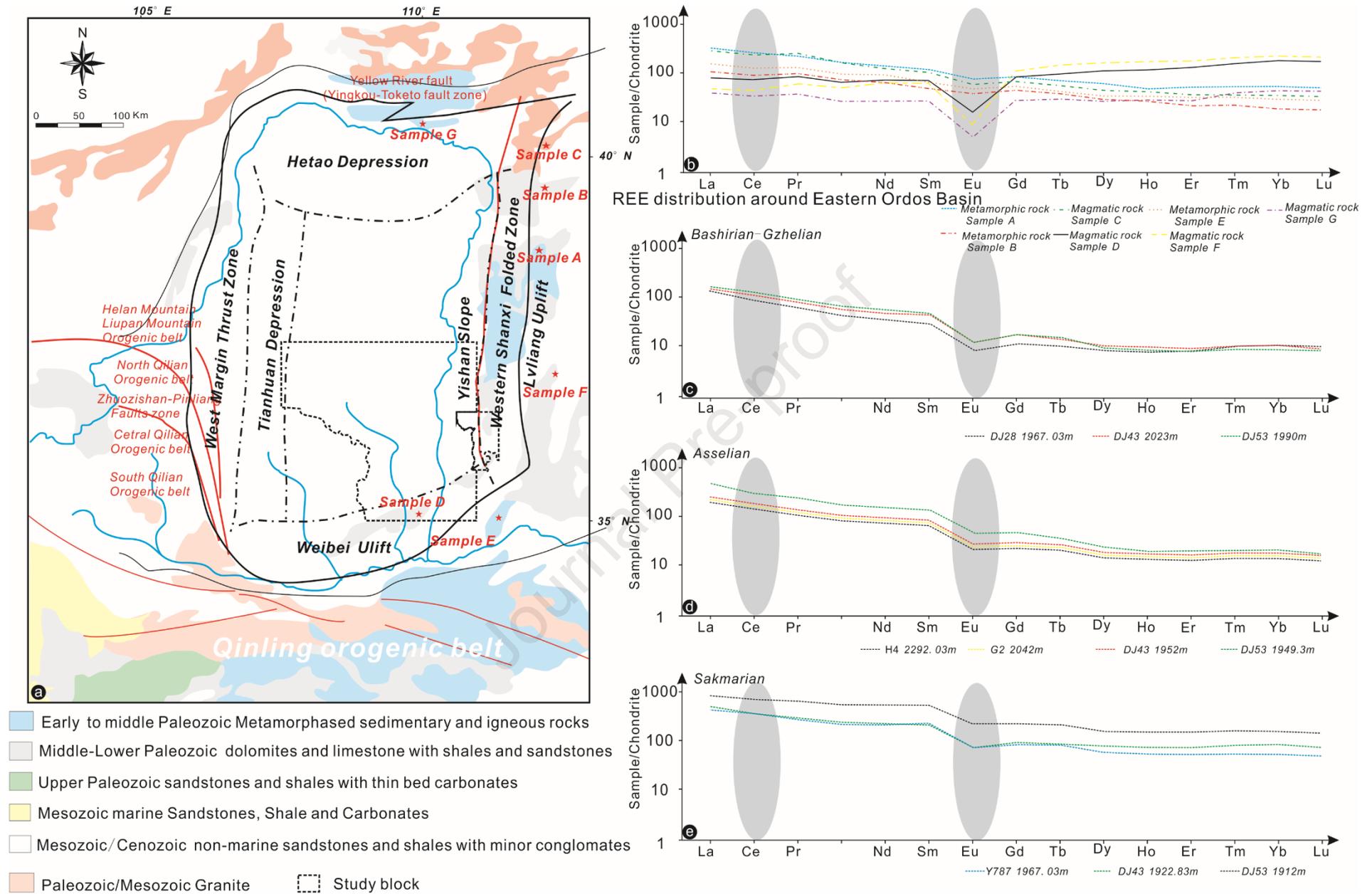


FIG 12

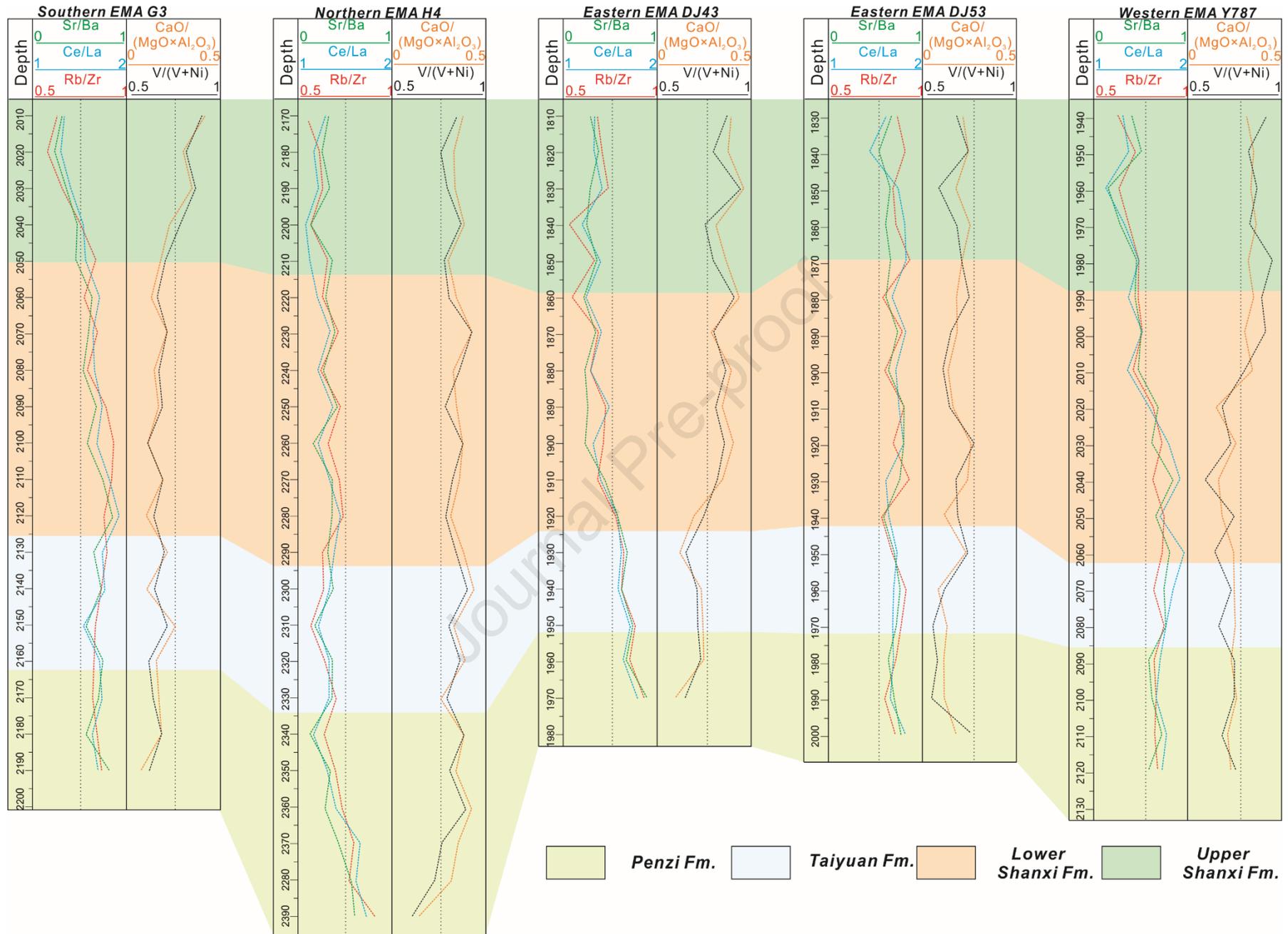


FIG 13

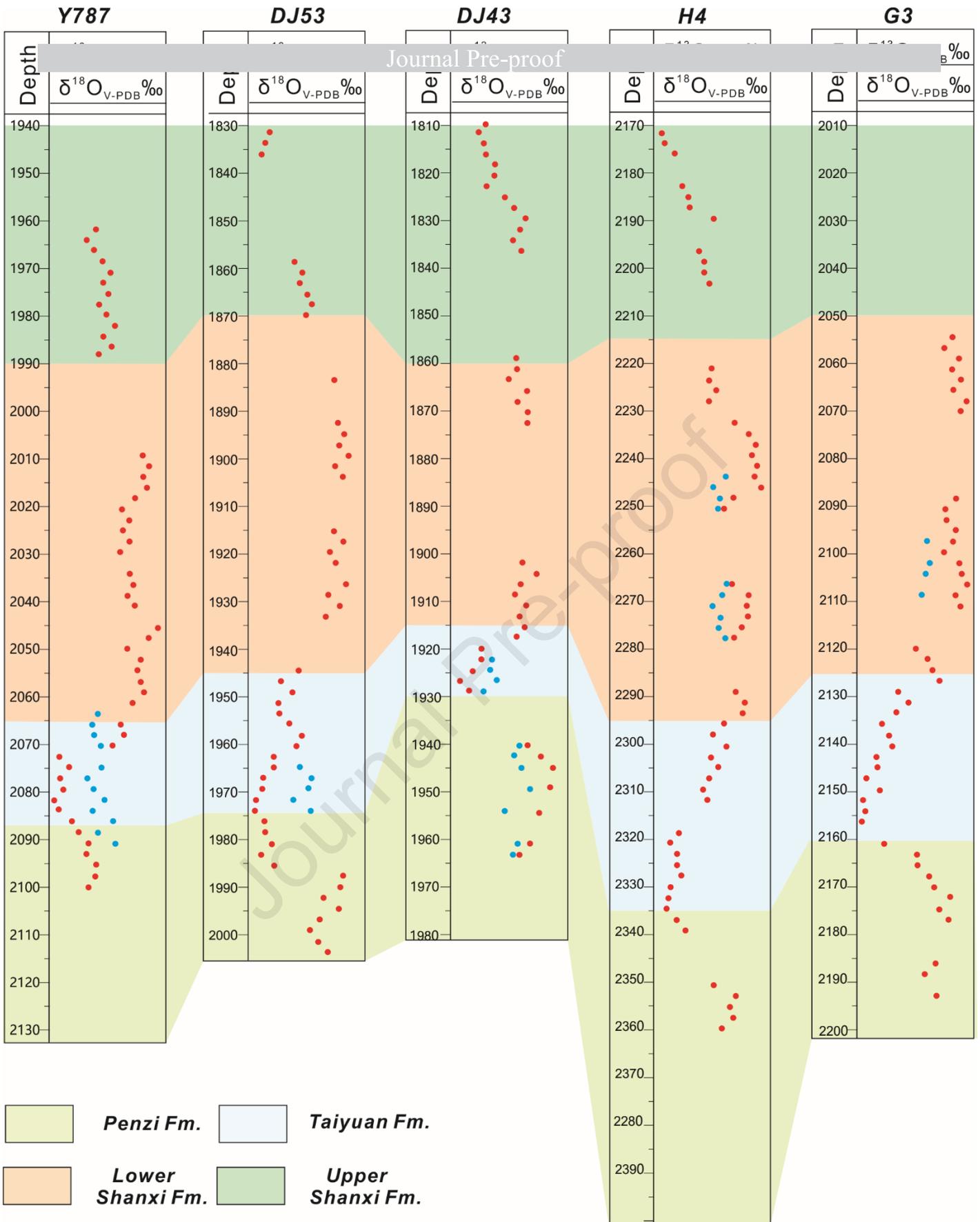
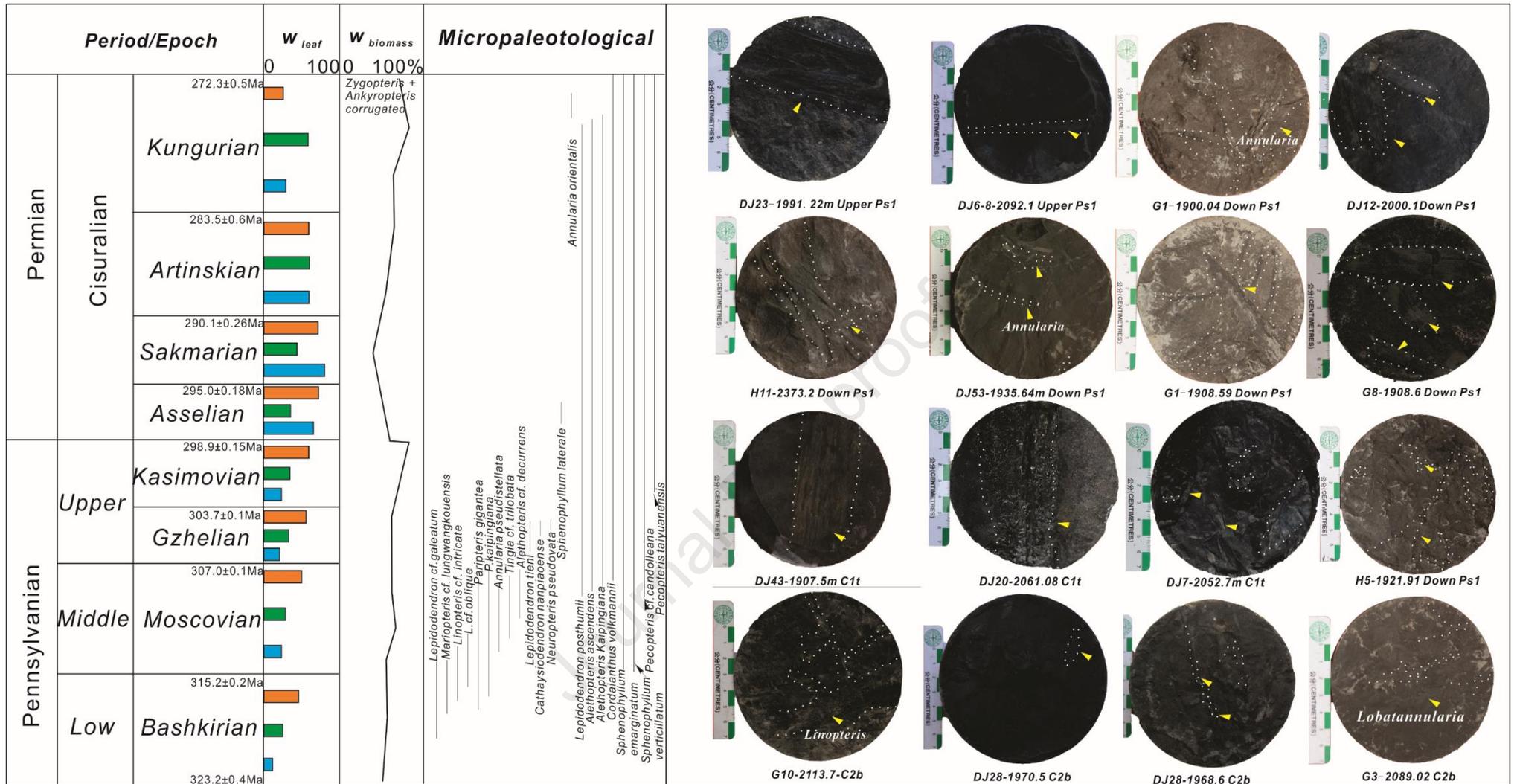
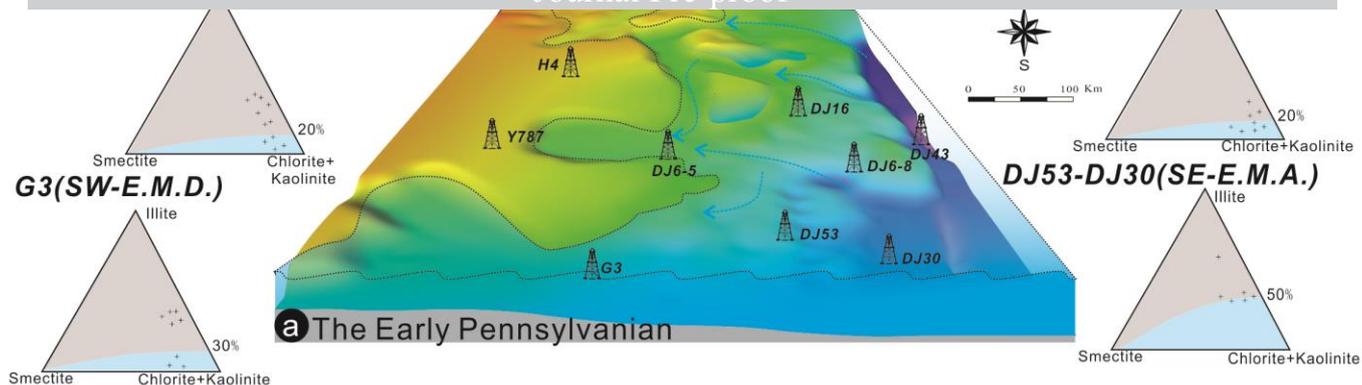


FIG 14



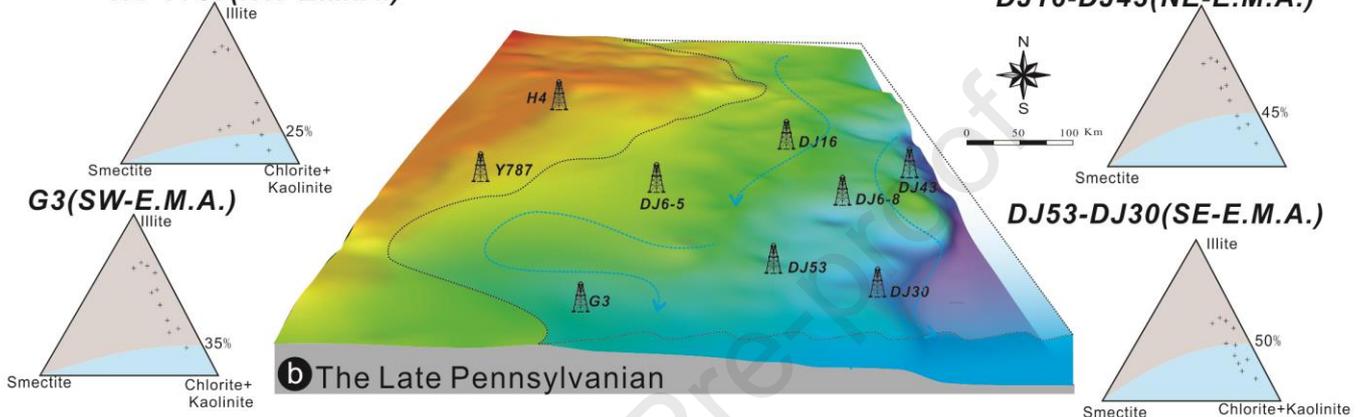
Root fossil
  Stem fossil
  Leaf fossil

FIG 15



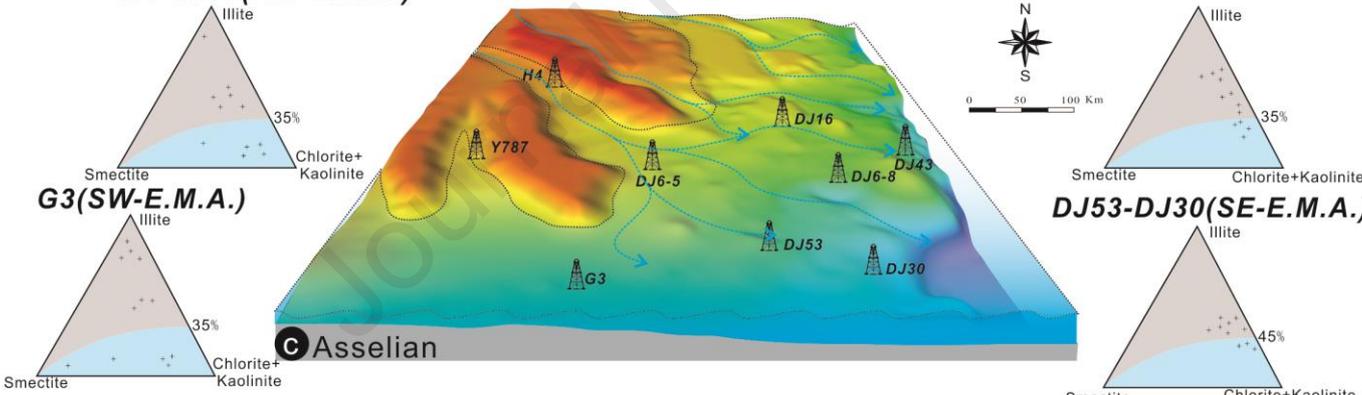
H4-Y787(NW-E.M.A.)

DJ16-DJ43(NE-E.M.A.)



H4-Y787(NW-E.M.A.)

DJ16-DJ43(NE-E.M.A.)



H4-Y787(NW-E.M.A.)

DJ16-DJ43(NE-E.M.A.)

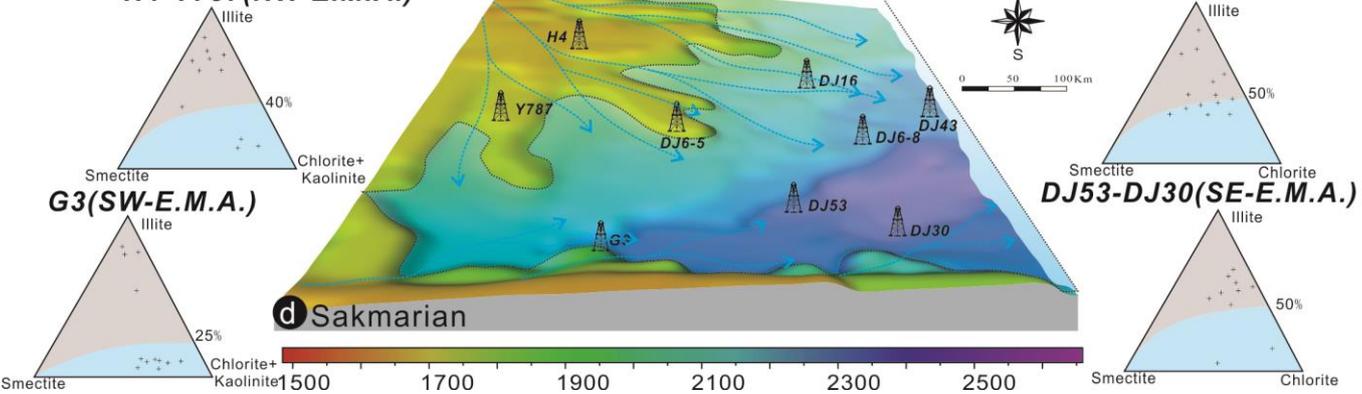


FIG 16

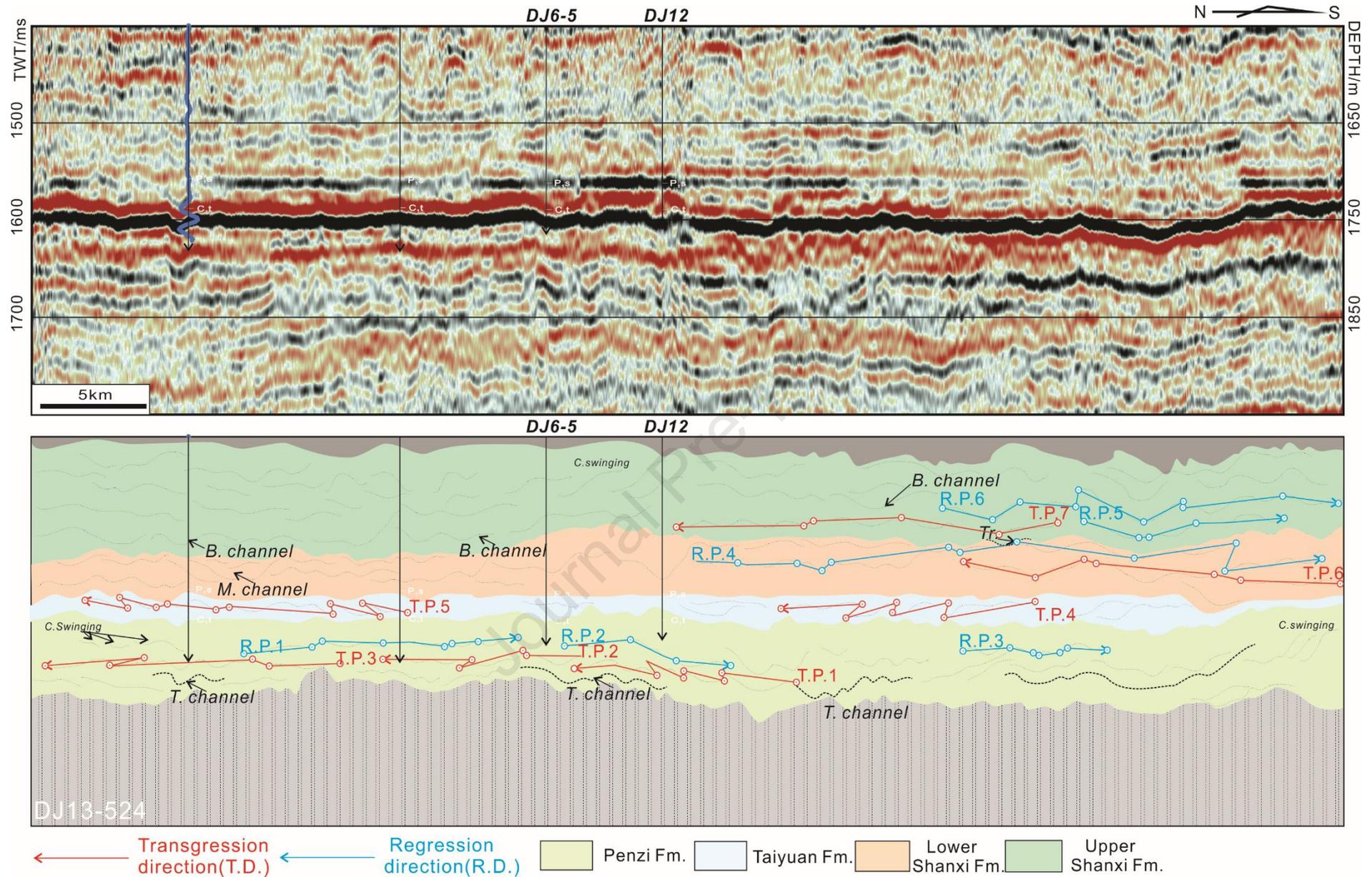
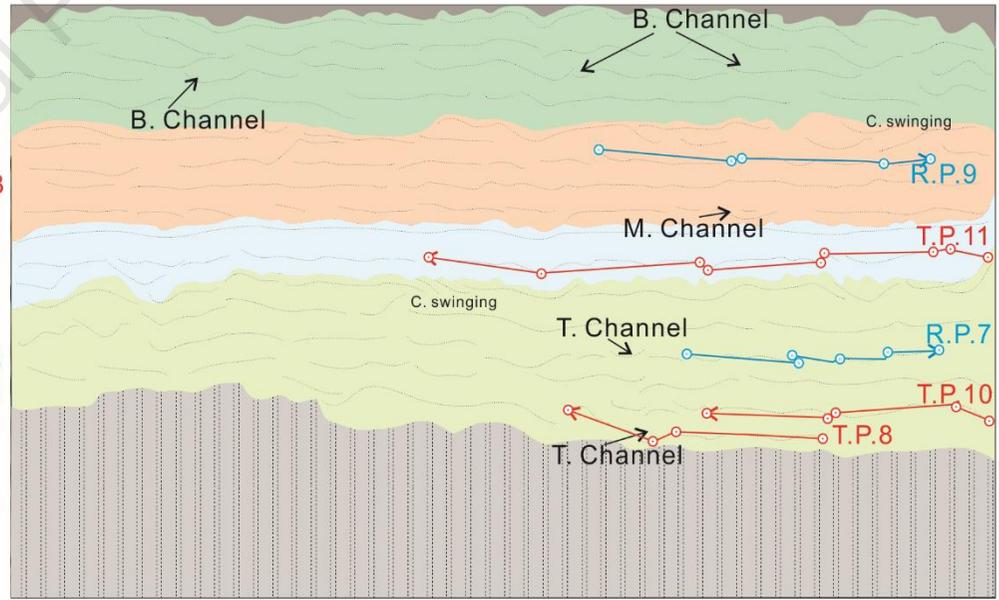
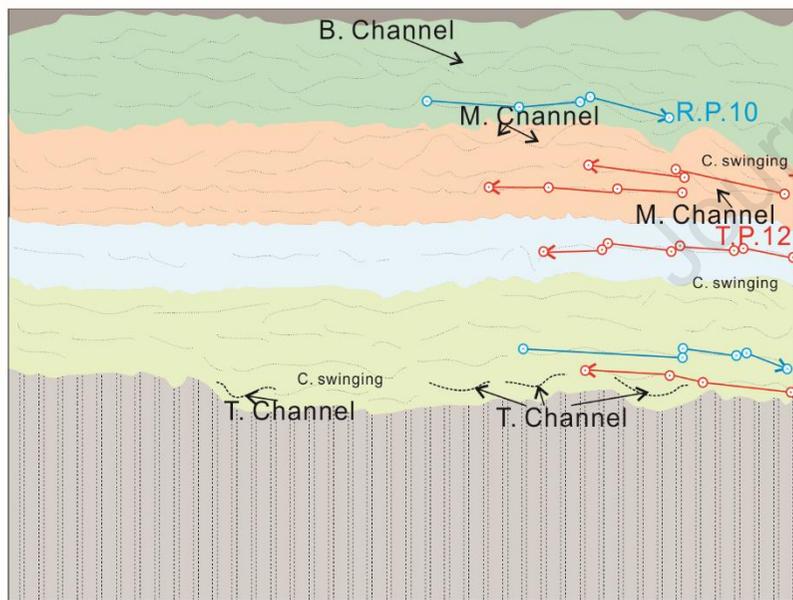
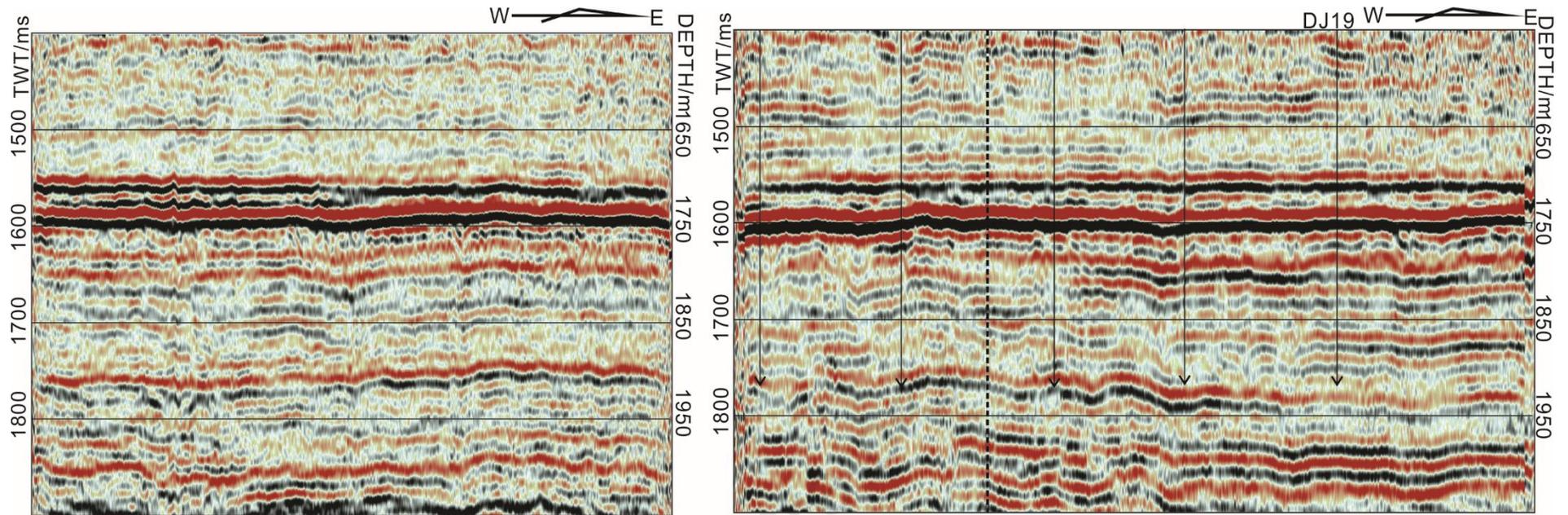


FIG 17



← Transgression direction (T.D.)     
 ← Regression direction (R.D.)

Penzi Fm.     
  Taiyuan Fm.     
  Lower Shanxi Fm.     
  Upper Shanxi Fm.

**FIG 18**

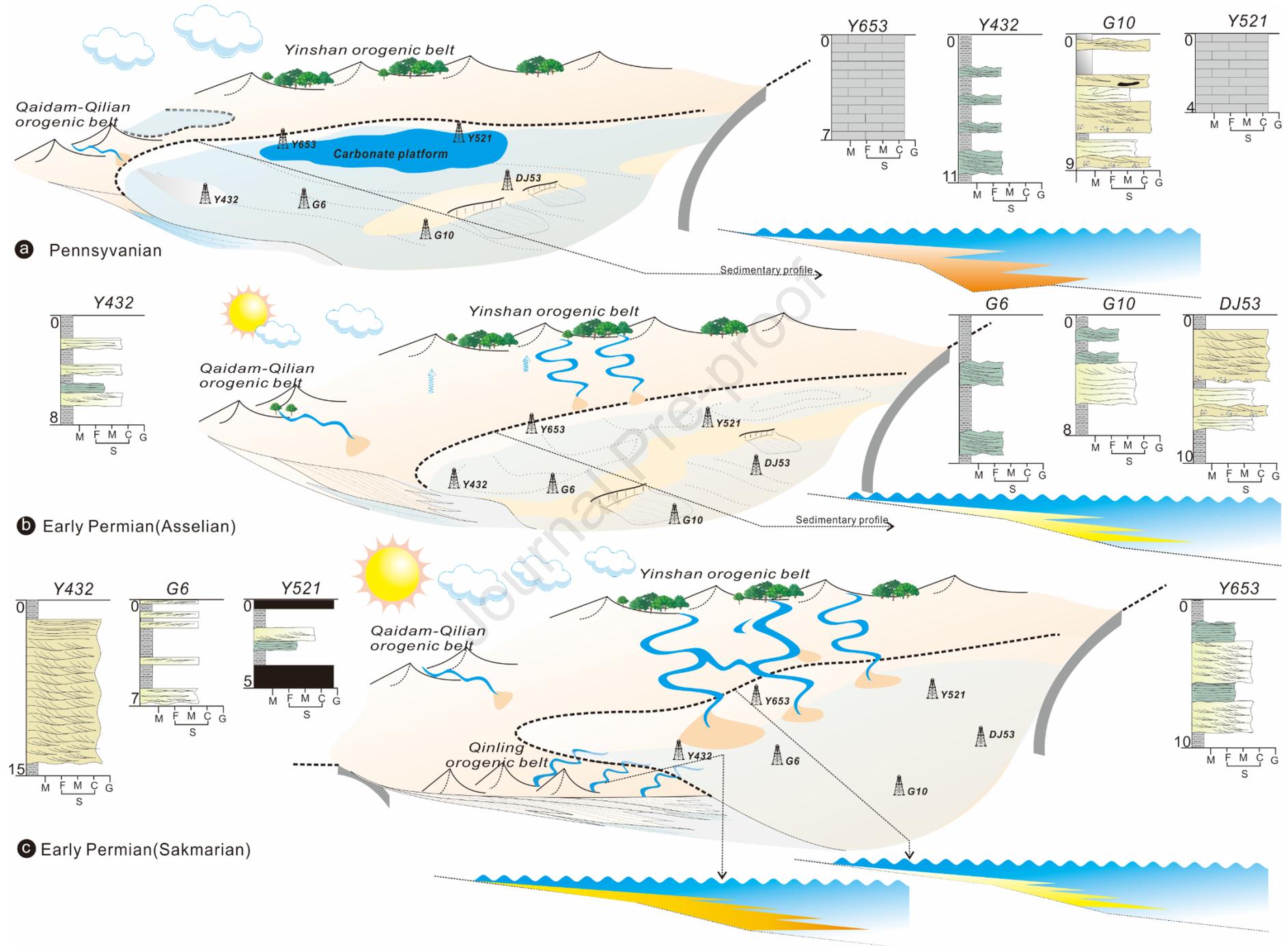


FIG 19

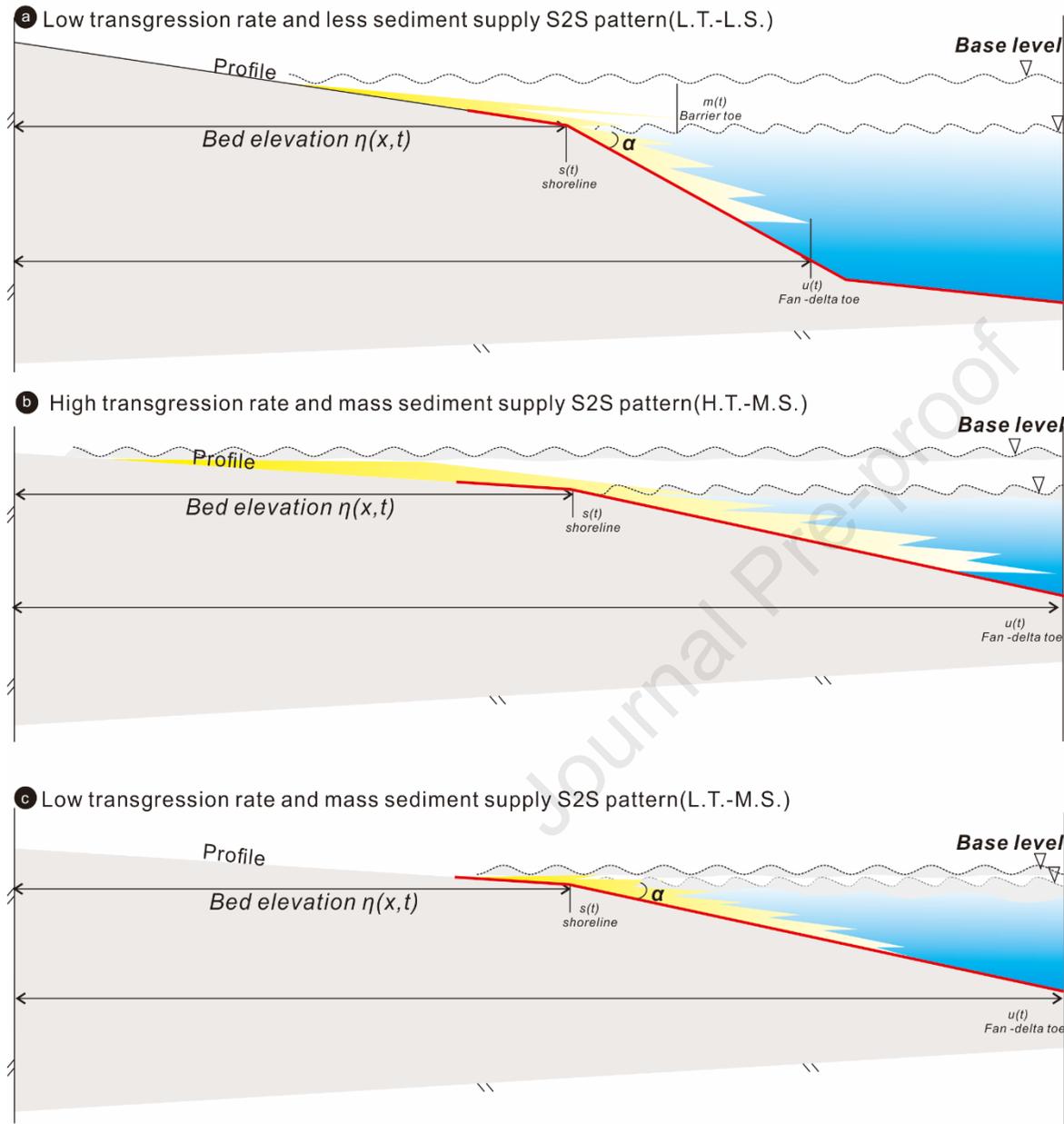
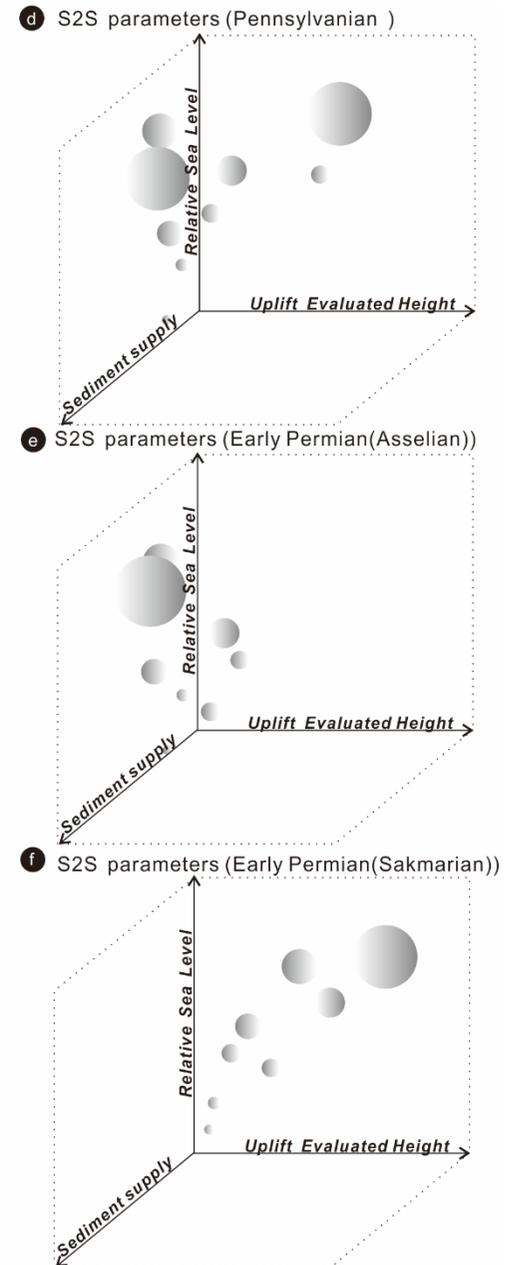


FIG 20



**Highlight**

1. Identified provenance area of the eastern margin of the Ordos Basin transfer from the Qaidam–Qilian orogenic belt to multiprovenance areas during the Pennsylvanian the late Carboniferous–early Permian.
2. Reconstructed the clockwise evolution process of the paleo sea level in the eastern margin of the Ordos Basin during the Paleozoic
3. Identified different types of source-to-sink systems insight from the spatial discrepancy of the sea level and sediment supply.
4. Proposed a integrated method to determine the dominating factors influencing the transgression/regression process

**Declaration of interests**

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.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

Journal Pre-proof