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1 **Controlling factors for differential subsidence in the Sonoma**
2 **Foreland Basin (Early Triassic, western USA)**

3
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20 Abstract –

21 Sediments deposited from the Permian–Triassic boundary (~252<5>Ma) until the end-Smithian (Early
22 Triassic; *c.* 250.7<5>Ma) in the Sonoma Foreland Basin show marked thickness variations between its
23 southern (up to *c.* 250<5>m thick) and northern (up to *c.* 550<5>m thick) parts. This basin formed as a
24 flexural response to the emplacement of the Golconda Allochthon during the Sonoma orogeny. Using a
25 high-resolution backstripping approach, a numerical model and sediment thickness to obtain a
26 quantitative subsidence analysis, we discuss the controlling factor(s) responsible for spatial variations
27 in thickness. We show that sedimentary overload is not sufficient to explain the significant discrepancy
28 observed in the sedimentary record of the basin. We argue that the inherited rheological properties of
29 the basement terranes and spatial heterogeneity of the allochthon are of paramount importance in
30 controlling the subsidence and thickness spatial distribution across the Sonoma Foreland Basin.

31 Keywords: Early Triassic, Sonoma orogeny, foreland basins, lithospheric strength, subsidence.

32 1. Introduction

33 The Sonoma Foreland Basin (SFB, western USA; Fig. 1a) provides an excellent Early
34 Triassic fossil and sedimentary record (Hofmann *et al.* 2014; Brayard *et al.* 2015; Thomazo *et al.*
35 2016). This N–S-trending foreland system (*sensu* DeCelles & Giles, 1996) was located on
36 the western Pangea margin and results from the emplacement of the Golconda Allochthon
37 (GA) during the Sonoma orogeny around the Permian–Triassic boundary (Fig. 1; Burchfiel &
38 Davis, 1975; Speed & Silberling, 1989; Ingersoll, 2008; Dickinson, 2013). Nevertheless,
39 despite numerous studies, the geometry and the palaeogeography of this basin remain poorly
40 constrained. The SFB covered a large area including present-day eastern Nevada, Utah, Idaho
41 and parts of Wyoming (Marzolf, 1993; Dickinson, 2006, 2013; Ingersoll, 2008).

42 Foreland sedimentary basins are generally considered as passive systems resulting from
43 the flexural subsidence of the elastic lithosphere in response to crustal thickening and
44 sediment loading (e.g. DeCelles & Giles, 1996; Allen & Allen, 2005). If the flexural isostatic
45 model is a reasonable first-order explanation for the overall shape of foreland basins,
46 sediment thickness variations and peculiar stratigraphic successions involve a differential
47 local subsidence. In order to decipher such potential mechanisms at the origin of the SFB
48 structuring and sedimentary record variations, we use a multidisciplinary approach. We
49 perform a subsidence analysis of the basin within a high-resolution biostratigraphically
50 controlled timeframe from the Permian–Triassic unconformity (PTU) up until late Smithian
51 time (a *c.* 1.3<5>Ma long interval; the Smithian is the third substage of the Early Triassic).
52 This allows us to characterize the basin infill in relation to the emplacement of the Golconda

53 Allochthon during the Sonoma orogeny. We also provide new evidence indicating that the
54 studied area is a foreland basin. Using a complementary backstripping approach and
55 numerical models we discuss the main factors controlling the subsidence variations observed
56 in the SFB, including the impact of lithospheric and rheological features, on basement
57 partitioning and sedimentation.

58 2. Geological setting

59 2.a. Brief geological history of the study area

60 The Sonoma Foreland Basin lies within a region of the North American continent
61 showing a very long and complex tectonosedimentary history starting during Proterozoic
62 time and still active today (e.g. Dickinson, 2013). The first documented structuring of the
63 region dates back to the Palaeoproterozoic period when Mojave and Yavapai terranes were
64 emplaced against the Archean Wyoming craton (Fig. 1b; Whitmeyer & Karlstrom, 2007, fig.
65 9; Lund *et al.* 2015). This event generated multiple crustal fault zones along which later
66 reactivations were possible with deformational episodes (Oldow *et al.* 1989; Dickerson,
67 2003). At least two rifting events took place in this region during subsequent Proterozoic
68 times (Burchfiel & Davis, 1975; Oldow *et al.* 1989), the most recent being during
69 Neoproterozoic time (*c.* 770–500 Ma) and linked to the fragmentation of the supercontinent
70 Rodinia (Fig. 1b; Dickinson, 2006). The long period of tectonic quiescence following the
71 formation of this passive margin lasted until Late Devonian time (*c.* 380–360 Ma) and
72 corresponds to the deposition of a thick sedimentary prism formerly known as the
73 ‘Cordilleran Miogeocline’ (Clark, 1957; Paull & Paull, 1991; Dickinson, 2006, 2013).

74 Starting during Late Devonian time and lasting until late Early Carboniferous time, the
75 Antler orogeny marks the beginning of a period of nearly continuous structural events that are
76 still active today (Fig. 1b). The Antler orogeny was caused by the convergence and accretion
77 of exotic island-arcs against the western margin of the North American Plate. This orogeny is
78 characterized by the emplacement of a large obducted accretionary prism located in Central
79 Nevada today (i.e. Roberts Mountains Thrust, Fig. 1c; Burchfiel & Davis, 1975; Speed &
80 Sleep, 1982; Speed & Silberling, 1989; Burchfiel & Royden, 1991). The Roberts Mountains
81 Allochthon led to the formation of the N–S-trending westwards-dipping Antler Foreland

82 Basin (Speed & Sleep, 1982; Burchfiel & Royden, 1991; Blakey, 2008; Ingersoll, 2008;
83 Dickinson, 2004, 2006, 2013).

84 Soon after the Antler orogeny the Ancestral Rocky Mountains (ARM) orogeny
85 occurred on the eastern part of the region (Fig. 1c), ranging over Early Carboniferous to
86 early–middle Permian time (*c.* 350–270 Ma; Fig. 1b). This mountain-building event
87 resulted from a succession of crustal uplifts because of important long-range intracratonic
88 deformations. There, transtensional and transpressional constraints occurred along with
89 lithospheric buckling as a response to the Laurentia–Gondwana continental collision (Kluth
90 & Coney, 1981; Ye *et al.* 1996; Geslin, 1998; Dickerson, 2003; Dickinson, 2006, 2013;
91 Blakey, 2008). The resulting chain probably showed a marked topographic relief, some of
92 which could have persisted until Early Triassic time (Kluth & Coney, 1981; Blakey, 2008).
93 Most of these crustal uplifts were emplaced according to lithospheric weaknesses inherited
94 from the Proterozoic structural events (Kluth & Coney, 1981; Dickerson, 2003).

95 Many sedimentary basins formed during the Carboniferous–Permian interval
96 (Dickerson, 2003). For instance, the Permian Oquirrh Basin (Fig. 1c) probably resulted from
97 the complex interplay between intracratonic deformations to the east and the reactivation of
98 Antler faults to the west (Geslin 1998: fig. 12; Trexler & Nitchman, 1990; Dickerson, 2003;
99 Blakey, 2008). This highly subsiding basin recorded up to 6 km of marine strata (Walker,
100 1985; Yonkee & Weil, 2015).

101 Similarly to the Antler orogeny, the Sonoma orogeny is the result of the eastwards
102 migration and accretion of exotic island-arc systems belonging to the Sonomia microplate
103 onto the North American Plate around the Permian–Triassic boundary (Burchfiel & Davis,
104 1975; Speed & Silberling, 1989; Dickinson, 2006, 2013; Blakey, 2008; Ingersoll, 2008). The
105 Sonoma orogeny is characterized by the thrusting of an accretionary prism above continental
106 crust, known as the Golconda Allochthon, and emplaced in the same area as the older Roberts
107 Mountains Allochthon (Fig. 1c). The Golconda Allochthon is thought to have initiated the
108 formation of a foreland basin – the Sonoma Foreland Basin (Dickinson, 2006, 2013; Blakey,
109 2008; Ingersoll, 2008) – which recorded sediments deposited during Early Triassic time.
110 However, field evidence pointing towards the location and extension of the Golconda
111 Allochthon is restricted to only a few remnants (e.g. ‘Koipato volcanics’) near the southern

112 part of the basin, which are presently located in Central Nevada (Fig. 1c; Snyder &
113 Brueckner, 1983; Walker, 1985; Schweickert & Lahren, 1987; Oldow *et al.* 1989; Dickinson,
114 2006, 2013; Blakey, 2008; Ingersoll, 2008). Remnants of the Golconda Allochthon, if any,
115 are yet to be found in the northern part of the basin, especially in Idaho (Schweickert &
116 Lahren, 1987; Oldow *et al.* 1989). This allochthon is sealed in present-day Nevada by the
117 rhyolitic Koipato Formation volcanism, presumably emplaced by the end of the Sonoma
118 orogeny (Vetz, 2011). A minimum age of Anisian (Middle Triassic) can be given to this
119 volcanic formation using geochronology (Vetz, 2011) and due to the occurrence of Anisian
120 ammonites in the unconformably overlying sedimentary series (Nichols & Silberling, 1977;
121 Bucher, 1988; Vetz, 2011). The potential presence of older ammonoid faunas is not to be
122 discarded.

123 The following Sevier orogeny is of Early Cretaceous – Eocene age (*c.* 140–50 Ma;
124 Fig. 1b) and it originated from the subduction of the Farallon Plate under the North American
125 continental plate (Burchfiel & Davis, 1975; Dickinson, 2006, 2013). E–W-directed
126 compressive constraints resulted in the formation of a large Sevier thrust-and-fold belt which
127 is still present today and constitutes the eastern border of the Great Basin (Fig. 1c; Dickinson,
128 2006, 2013; Yonkee & Weil, 2010; Yonkee *et al.* 2014). This thrust-and-fold belt is however
129 not homogeneous along its N–S-trending front, and displays two convex-to-the-foreland
130 ‘salients’ (Fig. 2) with varying estimated tectonic shortening and eastwards displacement of
131 terrains reaching up to 140 km (DeCelles & Coogan, 2006; Schelling *et al.* 2007;
132 Dickinson, 2006, 2013; Yonkee & Weil, 2010, 2015; Yonkee *et al.* 2014). These Wyoming
133 and Central Utah salients are separated by a conspicuous recess formed by a lateral ramp and
134 located west of the Uinta Mountains (Figs 1c, 2). Its formation results from inherited features
135 of the basement (see Section 4.c; e.g. Lawton, Boyer & Schmitt, 1994; Mukul & Mitra, 1998;
136 Paulsen & Marshak, 1999; Wilkerson, Apotria & Farid, 2002).

137 Also during Early Cretaceous – Eocene time, the eastern Laramide orogeny reactivated
138 basal crustal uplifts set during the Ancestral Rocky Mountains orogeny. This led to the
139 formation of the modern-day Rocky Mountains which overlapped older structures in the
140 Colorado Plateau (Fig. 1b, c; Oldow *et al.* 1989; Ye *et al.* 1996; Dickinson, pers. comm.
141 2015).

142 Finally, the Basin and Range extension of the entire region started during Neogene time
143 (c. 20<5>Ma; Fig. 1b) and is still active today (Oldow *et al.* 1989; DeCelles & Coogan, 2006;
144 Dickinson, 2002, 2006, 2013). This extension is the result of internal forces (Kreemer &
145 Hammond, 2007) that generated transtensional stresses and pure shear (Parsons, Thompson
146 & Sleep, 1994; Gans & Bohrsen, 1998; Dickinson, 2002, 2006). However, the origin of these
147 extensional constraints is still being discussed. Several possible mechanisms have been
148 proposed, including: (1) a mantellic ‘wide rift-like’ process with ascent and underplating of
149 mantellic material leading to thermal lamination of the lithosphere (Lachenbruch & Morgan,
150 1990; Parsons, Thompson & Sleep, 1994, Gans & Bohrsen, 1998); or (2) a mechanical origin
151 with the extension occurring in a late orogenic context, due to the instability and gravity
152 collapse of the thickened lithospheric crust present in Nevada and westernmost Utah
153 (Fletcher & Hallet 1983; Malavieille, 1993; Zandt, Myers & Wallace, 1995). Nevertheless,
154 the easternmost borders of the basin (e.g. Colorado Plateau or Uinta Mountains) are not
155 affected by these displacements (Fig. 1c; Dickinson, 2006, 2013). It is also worth noting that
156 this extension reactivates in inversion some of the thrust faults created during the Sevier
157 orogeny (Coney, 1987; Dickinson, 2006, 2013).

158 2.b. Sedimentary record of the Sonoma Foreland Basin

159 Here we focus on the Early Triassic sedimentary record of the Sonoma Foreland Basin
160 (Figs 3a, 4). The stratigraphic succession displays marked spatial differences in thickness and
161 in dominant lithologies (Fig. 4). The sedimentary record is considered as almost continuous
162 throughout the basin, with local erosion surfaces being under the temporal resolution of
163 ammonoid biozones for this Early Triassic interval (e.g. Olivier *et al.* 2014, 2016; Vennin *et*
164 *al.* 2015). In its southern part (Figs 3a, 4), the basin is mainly filled with transitional
165 continental to marine coarse sandstones to conglomerates known as ‘red beds’ of the
166 Moenkopi Group (Fig. 5a–c, e; *sensu* Lucas, Krainer & Milner, 2007; Brayard *et al.* 2013).
167 At the top of the Moenkopi Group, metric-scale beds of intertidal microbial limestones can be
168 observed (Figs 3a, 4, 5e; Brayard *et al.* 2013; Vennin *et al.* 2015; Olivier *et al.* 2016). The
169 upper part of the sedimentary pile is characterized by open-marine bioclastic limestones
170 (locally shales) of the Thaynes Group (Figs 3a, 4, 5d, f; *sensu* Lucas, Krainer & Milner,
171 2007), marking the maximum flooding of the Smithian third-order transgression (Embry,
172 1997; Vennin *et al.* 2015). This flooding event is characterized by the presence of the

173 ammonoid genus *Anasibirites* (Figs 3a, 4; Lucas, Krainer & Milner, 2007; Brayard *et al.*
174 2013; Jattiot *et al.* 2015, in press). In the northern part of the basin (Figs 3a, 4) the
175 sedimentary record differs at its base by the presence of the Dinwoody and Woodside
176 formations, characterized by fine marine siltstones (Figs 3a, 4, 5g; Kummel, 1954, 1957;
177 Sadler, 1981; Paull & Paull, 1991). Above these formations, the sedimentary record
178 resembles that observed in the southern part and corresponds to the open-marine bioclastic
179 limestones and shales of the Thaynes Group (Figs 3a, 4, 5d, h). A basin-scale synthetic facies
180 analysis with associated depositional environments and estimations of the palaeobathymetries
181 can be found in online Supplementary Table S1 (available at
182 <http://journals.cambridge.org/geo>).

183 3. Dataset and methods

184 3.a. Dataset

185 We compiled a comprehensive sedimentary and biostratigraphic dataset for the Early
186 Triassic outcrops in the Sonoma Foreland Basin, including previously published works (e.g.
187 Kummel, 1954, 1957; Paull & Paull, 1991; Goodspeed & Lucas, 2007; Heckert *et al.* 2015)
188 together with new field data (Fig. 3b). We selected 43 biostratigraphically correlated sections
189 documenting different parts of the basin in order to estimate the thickness (at the metre scale)
190 of the sedimentary deposits (GPS coordinates and main characteristics of each section are
191 provided in online Supplementary Table S2). The 43 studied sections correspond to the Early
192 Triassic interval. The base of this interval is defined by a major regional PTU (Brayard *et al.*
193 2013). Its upper end is determined by the *Anasibirites* beds or the uppermost part of the
194 *Owenites* beds as a surrogate, which are the main biostratigraphic markers of the end-
195 Smithian (Figs 3a, 4; Brayard *et al.* 2013; Jattiot *et al.* 2015). Eleven sections were delimited
196 using a high-resolution ammonoid zonation (e.g. sections in Fig. 4; Brayard *et al.* 2013). We
197 conservatively used only minimum thickness values for the 32 sections taken from the
198 literature because they are not always based on homogeneous sedimentary and
199 biostratigraphical data (online Supplementary Table S2). For completeness of the subsidence
200 analysis, we included when possible thickness data available for the lower part of the
201 Spathian (fourth substage of the Early Triassic), the *Columbites* beds marking in this case the
202 end of the studied interval (Fig. 3a).

203 3.b. Methods

204 3.b.1. Palinspastic reconstructions using retrodeformations

205 Post-Triassic times in the Sonoma Foreland Basin are characterized by important
206 tectonic compressive and later extensive deformations. These successive deformations are
207 mostly represented in the basin by the complex and heterogeneous Sevier thrust-and-fold
208 belt. The palaeogeographic configuration of the Sonoma Foreland Basin was therefore
209 different compared to the modern configuration. In order to resolve this issue, we performed
210 a palinspastic reconstruction to estimate the Early Triassic palaeogeography of this basin.

211 Retrodeformations of observed structural features affecting the Triassic series were
212 applied to several regional cross-sections using literature data (e.g. DeCelles & Coogan,
213 2006; Yonkee & Weil, 2010; Fig. 6). This method consists of the horizontalization of a
214 selected layer (here the Triassic series) by virtually inverting all the structural features
215 observed in the section between a fixed reference point named the ‘pin line’ and a mobile
216 reference point named the ‘loose line’ (Fig. 6; see Groshong, 2006 for details). In the two
217 regional cross-sections of the Sevier thrust-and-fold belt illustrated in Figure 6, most
218 structural features are thrust complexes; horizontalization therefore mainly consists of
219 retrodeformation of the displacements along thrust planes. Finally, balanced cross-sections
220 represent a good approximation of the geomorphological setting by the time of deposition.
221 Based on this method, the direction and value of the estimated tectonic transport (ETT)
222 underwent by the terrains can also be calculated (e.g. *c.* 140<5>km and *c.* 60<5>km for the
223 cross-sections a and b in Fig. 6, respectively).

224 Due to the complex nature of the Sevier thrust-and-fold belt resulting from the inherited
225 structure and thickness pattern of the pre-deformation basins (Paulsen & Marshak, 1999), and
226 also the westwards focalization of the subsequent Basin and Range extension, ETT was
227 spatially heterogeneous between Wyoming and Central Utah salients (Mukul & Mitra, 1998;
228 DeCelles & Coogan, 2006; Schelling *et al.* 2007; Yonkee & Weil, 2010; Yonkee *et al.* 2014).
229 We therefore defined seven sectors within our study area (sectors 1–7 in Fig. 7). These
230 sectors were delimited based on similar ETT values (Table 1; Fig. 7). These values were
231 determined from data available in the literature (references in Table 1) and checked with the

232 retrodeformation of regional cross-sections taken from geological maps (cross-sections in
233 Fig. 6).

234 3.b.2. Subsidence analysis and backstripping

235 Subsidence analysis quantifies the vertical movements underwent by a given
236 sedimentary depositional surface through a graphic representation, by tracking the subsidence
237 and uplift history of said surface (Van Hinte, 1978). This history is reconstructed based on
238 sedimentary thickness, lithology, palaeo-sea level, palaeobathymetry and age data. This
239 analysis also accounts for the mechanical compaction underwent by the sediments. The
240 resulting curve provides a view of the total subsidence history for a given stratigraphic
241 column (Van Hinte, 1978; Allen & Allen, 2005). Steckler & Watts (1978) showed that the
242 local isostatic effect exerted by the sedimentary load can be removed. This ‘backstripping’
243 method can therefore help to characterize the tectonic subsidence only, as if the basin has
244 been filled by air only and not by water and/or sediment during its history (Steckler & Watts,
245 1978; Xie & Heller, 2009). Backstripping is also used to restore the initial thickness of a
246 sedimentary column (Angevine, Heller & Paola, 1990; Allen & Allen, 2005). Lithological
247 compositions and palaeobathymetries have been checked using facies analysis (online
248 Supplementary Table S1) or literature data (see analysed sections in Fig. 3b and online
249 Supplementary Table S2). Porosity was quantified by comparison with experimental data
250 (e.g. Van Hinte, 1978; Sclater & Christie, 1980) and represents an important proxy for
251 compaction analysis. Additionally, Chevalier *et al.* (2003) and Lachkar *et al.* (2009) showed
252 that a highly resolved biostratigraphic control is useful to define and quantify variations in
253 subsidence at a fine spatio-temporal scale as it yields accurate subsidence rates. For the Early
254 Triassic Sonoma Foreland Basin, the high-resolution ammonoid zonation by Brayard *et al.*
255 (2013) serves as the main timeframe. Complementary absolute time lines were obtained from
256 radiometric ages published from coeval beds in South China (Galfetti *et al.* 2007; Burgess,
257 Bowring & Shen, 2014), whereas the duration of the studied intervals was interpolated from
258 ammonoid biozone duration (after Brühwiler *et al.* 2010 and Ware *et al.* 2015). Palaeo-sea
259 level curve is based on data from Haq, Hardenbol & Vail (1988), providing a quantitative
260 representation of the reconstructed Early Triassic sea level.

261 We chose to not use the flexural backstripping method (Allen & Allen, 2005) due to the
262 lack of appropriate data needed for such model (e.g. flexural rigidity data, regional
263 distribution of the sedimentary load). Instead, we calculated the total and tectonic subsidence
264 curves using the one-dimensional (1D) local isostatic approach of Steckler & Watts (1978).
265 In addition, this method emphasizes the tectonic subsidence as ‘a way of normalizing
266 subsidence in different basins that have undergone very different sedimentation histories’
267 (Xie & Heller, 2009). Our results for the tectonic subsidence history in the SFB can therefore
268 be compared to the compilation of Xie & Heller (2009). Subsidence analyses were performed
269 on four sections (Fig. 3b) using the OSXBackstrip software performing 1D Airy
270 backstripping (after Watts, 2001; Allen & Allen, 2005; available at:
271 <http://www.ux.uis.no/~nestor/work/programs.html>). These sections were selected for their
272 completeness (a complete and continuous sedimentary succession is reported from the PTU
273 to at least lower Spathian stratigraphy), for the presence of biostratigraphic markers
274 (ammonoid beds) and for their repartition within the SFB (representative of both the northern
275 and southern areas). A complete set of initial parameters and detailed results of the
276 subsidence analysis for each of the four sections are reported in online Supplementary
277 Material S1.

278 This analysis bears limitations as some errors may arise from uncertainties around the
279 data used for the subsidence analysis (Chevalier *et al.* 2003; Xie & Heller, 2009): (1)
280 accuracy of the measurement and report of the sedimentary thickness; (2) backstripping
281 calculation; (3) palaeo-bathymetry estimations; and (4) age control. Regarding the accuracy
282 of the sediment thickness, all selected sections have been measured at a centimetric scale.
283 Errors on measurements are therefore rather low, i.e. $\pm 2\%$ of the total thickness. In the
284 backstripping analysis, variables used for the calculation of the burial compaction are:
285 thickness; the initial porosity of the sediment; and the lithological constant of corresponding
286 lithologies. The latter two parameters are determined by comparison with experimental data
287 (e.g. Van Hinte, 1978; Sclater & Christie, 1980). Error on sediment decompaction is therefore
288 estimated to be low (*c.* $\pm 5\%$). Palaeobathymetry is hard to determine because of the
289 paucity of discriminating indicators. We hypothesize that errors on depth estimations are
290 about $\pm 10\%$. For age control, we used a compilation of biostratigraphic and

291 radiochronological data, leading to a detailed timeframe with a maximum error of around
292 60<5>ka (Brühwiler *et al.* 2010).

293 3.b.3. Spatial distribution of sedimentary thickness

294 PTU-Smithian sedimentary thicknesses and their respective location within the SFB
295 were integrated in Global Mapper v.16.2.3 GIS software (available at
296 <http://www.bluemarblegeo.com/products/global-mapper.php>) to generate an isopach map by
297 creating a 3D triangulated grid projection of thicknesses (online Supplementary Figure S1).

298 3.b.4. Lithospheric heterogeneity of the basement

299 To explore the nature of the SFB basement, a terrane map was constructed using
300 previous published maps by Yonkee *et al.* (2014), Yonkee & Weil (2015) and Lund *et al.*
301 (2015). In addition, we analysed several types of geophysical data: a raw regional Bouguer
302 gravity anomaly map (Kucks, 1999); an aeromagnetic anomaly map from Bankey *et al.*
303 (2002); and literature data (e.g. Gilbert, Velasco & Zandt, 2007). We also used published
304 U/Pb radiochronological data to assess an age for each basement terrane defined in the basin
305 (Foster *et al.* 2006; Fan *et al.* 2011; Mueller *et al.* 2011; Nelson, Hart & Frost, 2011;
306 Strickland, Miller & Wooden, 2011). It is worth noting that Precambrian crystalline
307 basements, lying under detachments and décollements responsible for nucleation of thrusting,
308 are not affected by these ‘thin-skin’ thrust-induced displacements (DeCelles & Coogan, 2006;
309 Schelling *et al.* 2007; Yonkee & Weil, 2010).

310 3.b.5. Numerical model

311 The flexural response of the SFB basement has been simulated using a 2D plane stress
312 flexural model solved with a finite element method code written in Matlab®. This approach
313 has been successfully used to model lithospheric deformation due to topographic and mantle
314 loads (Le Pourhiet & Saleeby, 2013) and ice loads (Moreau *et al.* 2015). First, a model of the
315 basin is made using field-based and literature data to characterize and quantify the flexural
316 response of the modelled SFB basement. Three additional models are then proposed to test
317 different scenarios regarding possible mechanisms controlling the flexure of the SFB
318 basement.

319 4. Results

320 We first reconstructed the SFB palaeogeography and used lithological and
321 stratigraphical analyses to constrain the spatial distribution of the sedimentary record across
322 the basin. This approach provides estimations of subsidence rates in the SFB. Secondly, we
323 identified and characterized the terranes that compose the SFB basement using geophysical
324 and cartographic data, as well as previously published ages. We then reconstructed the
325 morphology of the Golconda Allochthon in relation to the heritage of the basin. Finally, a 2D
326 model is proposed to quantify the flexural behaviour of the basin.

327 4.a. Lithological and stratigraphical analyses

328 Previous palaeogeographic reconstructions of the SFB did not take tectonic events and
329 the ensuing displacements into account (e.g. Paull & Paull, 1993). The palinspastic map of
330 the basin with the initial locations of the studied sections is shown on Figure 7. For the first
331 time post-Triassic displacements were accounted for, including: (1) the Sevier orogeny
332 (Cretaceous–Eocene) and the associated regional shortening due to the setting of a thrust-and-
333 fold belt (e.g. Yonkee & Weil, 2010); and (2) the later Neogene – present-day extension
334 linked to the Basin and Range province (e.g. Yonkee *et al.* 2014).

335 Based on the palinspastic map, we constructed a palaeogeographic isopach map of the
336 SFB (Fig. 8). The isopach map shows that the distribution of the sedimentary thickness for
337 the PTU-Smithian interval is heterogeneous within the basin, showing a thicker succession in
338 the northern than in the southern part. In the southern part, the thickness gradually varies
339 along a roughly NW–SE-aligned transect, showing low thicknesses over a large surface
340 (c. 500 km from east to west). The thickness ranges from a few tenths of metres in south
341 and SE Utah, up to 250 m around Salt Lake City. The westernmost area (NE Nevada) is
342 also characterized by low thicknesses (<100 m thick). Conversely, the northern part of the
343 basin exhibits a marked transition with thickness values broadly increasing from east to west.
344 The easternmost area of the northern part (west Wyoming) shows sedimentary thicknesses
345 similar to that of the southern part (<300 m thick; Fig. 8). The west-central area records
346 the thickest succession of the SFB (up to c. 550 m thick), and is centred on present-day
347 south-central Idaho. The westernmost area (west-central Idaho) shows similar thicknesses (up
348 to c. 300 m thick; Fig. 8).

349 The subsidence analysis (Fig. 9) also shows a clear distinction between the northern
350 and southern parts of the basin. Confusion Range (CR, Fig. 9a) and Pahvant Range (PR,
351 Fig. 9b) sections exhibit relatively low subsidence curves during the studied interval, whereas
352 Sheep Creek (SC, Fig. 9c) and Hot Springs (HS, Fig. 9d) sections show a high subsidence
353 profile. The total and tectonic subsidence curves are similar and the tectonic subsidence is
354 here a major component of the total subsidence, accounting for at least two-thirds of the total
355 subsidence, if not more (e.g. in CR, Fig. 9a).

356 When looking at the dominant lithologies (Fig. 9e), the sections from the southern part
357 of the basin display a sedimentary succession dominated by coarse conglomerates and
358 sandstones and microbial limestones of the Moenkopi Group and the limestones/shales of the
359 Thaynes Group (Figs 3, 4, 9e), while the total subsidence is low. By contrast, the sections
360 from the northern part of the SFB are dominated by fine siltstones (Figs 3, 4, 9e) with an
361 important subsidence.

362 Finally, the tectonic subsidence appears as a critical diagnostic feature for the basin
363 (Fig. 9f). A marked difference exists between mean tectonic subsidence rates in the southern
364 and northern parts of the basin (*c.* 100–500 m Ma^{-1} v. *c.* 500–600 m Ma^{-1} , respectively). The
365 southern sections show a low-rate tectonic subsidence (50–200 m Ma^{-1} ; Fig. 9e).
366 Nevertheless, a marked increase in subsidence rate is recorded during early Spathian time for
367 these sections (150–600 m Ma^{-1} ; Fig. 9e). Conversely, the northern sections show a
368 higher rate of tectonic subsidence during the PTU-Smithian interval (450–650 m Ma^{-1} ;
369 Fig. 9e), whereas early Spathian time is characterized by a decrease in subsidence rate (100–
370 250 m Ma^{-1} ; Fig. 9e).

371 4.b. Basement characterization

372 On the gravimetric anomaly map shown on Figure 10a, black lines outline the
373 geophysical features that may represent traces of crustal/lithospheric faults or heterogeneities
374 in the basement (Lowrie, 2007). The lowest Bouguer anomaly values (<150 mGal ,
375 Fig. 10a) suggest the presence of a thick crust, whereas moderate negative anomalies
376 (between –65 and –135 mGal ; white outlines) point towards a thinner crust and/or the
377 presence of lower-crustal high-density bodies (e.g. Gilbert, Velasco & Zandt, 2007; Lowrie,
378 2007). The Snake River Plain (SRP in Fig. 10a) is a Yellowstone hotspot track-related

379 basaltic province. This young (of Neogene age) structure influences neither the geometry nor
380 the properties of the basement (Dickinson, 2013). The Farmington Anomaly (FA on
381 Fig. 10a), located in the centre of the study area, may result from the presence of lower-
382 crustal high-density mafic and/or ultramafic material emplaced during a thermal event dated
383 at *c.* 1.64 Ga (Mueller *et al.* 2011). Alternatively, it can have originated from a more
384 recent thermal event and/or the presence of a thin lithospheric crust (e.g. Gilbert, Velasco &
385 Zandt, 2007; Lowrie, 2007). Remnants of an important thermal metamorphism including
386 partial melting (*c.* 1.67 Ga) can also be observed in this area (red dots in Fig. 10c; Mueller
387 *et al.* 2011). The Southern Anomaly (SA on Fig. 10a) is poorly documented and may result
388 from variations in the crustal thickness of the terrane (e.g. Gilbert, Velasco & Zandt, 2007;
389 Lowrie, 2007), possibly linked to the Ancestral Rocky Mountains orogeny or to the more
390 recent Laramide orogeny and the building of the Rocky Mountains (Ye *et al.* 1996;
391 Dickerson, 2003).

392 The aeromagnetic anomaly map presented in Figure 10b discriminates areas of
393 contrasted magnetic signatures (separated by black lines on Fig. 10b). These disturbances in
394 magnetic field are attributed to differences in the nature of the rocks composing the basement
395 (Turner, Rasson & Reeves, 2007). We do not attempt to identify the exact nature of these
396 rocks here; rather, we use these contrasted anomalies to characterize differences of rock types
397 that compose the basement (Purucker & Whaler, 2007; Lund *et al.* 2015). As on the Bouguer
398 gravity anomaly map, the presence of the Snake River Plane hotspot-track (SRP in Fig. 10a,
399 b) is obvious on the aeromagnetic anomaly map. It features a strong positive magnetic
400 anomaly signal (>150 nT, Fig. 10b). The Southern magnetic Zone (SZ on Fig. 10b) can be
401 distinguished on the southern part of the studied area by contrasted anomalies with a wide
402 range of variations (from *c.* -200 nT up to *c.* 400 nT). The Central magnetic Zone (CZ
403 on Fig. 10b) occupies the central third of the map. It is characterized by generally neutral to
404 (strongly) positive anomalies (from *c.* -10 nT to *c.* 60 nT, locally up to >150 nT).
405 In the northeastern quarter of the studied area, the North-Eastern magnetic Zone (NEZ on
406 Fig. 10b) is characterized by generally negative anomalies (between *c.* -80 nT and *c.* $-$
407 10 nT). Some areas with strong positive anomalies (>150 nT) are also observed,
408 whose shape and extension are very similar in the Bouguer gravity anomaly map (Fig. 10a).
409 Finally, a small Northern magnetic Zone (NZ on Fig. 10b) is visible north to the SRP and

410 west to the NZ. It shows contrasting anomalies, but with a less important range of variation
411 than the SRP and less strongly positive values (from *c.* $-60 <5>nT$ to *c.* $150 <5>nT$ only).

412 Figure 10c synthesizes the location and the different U/Pb radiochronological ages for
413 the basement (Foster *et al.* 2006; Fan *et al.* 2011; Mueller *et al.* 2011; Nelson, Hart & Frost,
414 2011; Strickland, Miller & Wooden, 2011). Basement rocks of Archean, Palaeoproterozoic
415 and Mesoproterozoic ages can be found throughout the entire studied area (Fig. 10c).
416 Archean ages are found in Wyoming, southwestern Montana and northeastern Nevada
417 (Fig. 10c; Fan *et al.* 2011; Mueller *et al.* 2011; Nelson, Hart & Frost, 2011; Strickland, Miller
418 & Wooden, 2011). Palaeoproterozoic ages are found in Utah and eastern Nevada (Fig. 10c;
419 Mueller *et al.* 2011; Nelson, Hart & Frost, 2011). Finally, Mesoproterozoic ages associated
420 with metamorphism are found in northwestern Utah and northern Idaho (Fig. 10c; Foster *et*
421 *al.* 2006; Mueller *et al.* 2011; Nelson, Hart & Frost, 2011).

422 Five different lithospheric terranes composing the SFB basement can therefore be
423 identified: the Wyoming Terrane (WT); the Grouse Creek Block (GCB); the Mojave Terrane
424 (MT); the Yavapai Terrane (YT); and the Farmington Terrane (FT; Fig. 10d). The GCB and
425 WT are Archean terranes with ages of *c.* $2.5 <5>Ga$ (Nelson, Hart & Frost, 2011; Strickland,
426 Miller & Wooden, 2011) and $2.4\text{--}3.3 <5>Ga$ (Fan *et al.* 2011; Mueller *et al.* 2011),
427 respectively. The MT is a Palaeoproterozoic terrane of age $2.04\text{--}2.34 <5>Ga$, whereas the YT
428 is a younger Palaeoproterozoic terrane of age $1.720\text{--}1.744 <5>Ga$ (Nelson, Hart & Frost,
429 2011). The FT is a Mesoproterozoic intracratonic mobile belt (Lund *et al.* 2015) composed of
430 reworked Archean crust (Whitmeyer & Karlstrom, 2007), with metamorphism ages between
431 1.63 and $1.71 <5>Ga$ (Foster *et al.* 2006; Mueller *et al.* 2011; Nelson, Hart & Frost, 2011).

432 4.c. Impact of the heritage on the SFB development

433 The fact that the basement of the SFB is composed of five Archean–Mesoproterozoic
434 terranes questions the potentially crucial role of inherited lithospheric features on the
435 formation and spatio-temporal evolution of the SFB.

436 Lithospheric strength (i.e. rigidity) of the terranes varies depending on their age and
437 heritage (Poudjom Djomani *et al.* 2001; Artemieva & Mooney, 2002), with important
438 changes in rheological behaviour and segregation between oldest ($>1.7 <5>Ga$) and juvenile

439 crusts (<1.7<5>Ga; Artemieva & Mooney, 2002). Since older lithospheres are more rigid
440 than younger, Archean and Palaeoproterozoic basements such as the Wyoming Terrane,
441 Grouse Creek Block, Mojave Terrane and Yavapai Terrane are defined here as ‘strong’
442 lithospheres (e.g. Cardozo & Jordan, 2001; Leever *et al.* 2006; Fig. 11). Conversely, the more
443 recent Mesoproterozoic lithospheres such as the Farmington Terrane (Fig. 11) are
444 characterized by a lower rigidity (e.g. Cardozo & Jordan, 2001; Leever *et al.* 2006; Fosdick,
445 Graham & Hilley, 2014). Additionally, some lithospheres can be weaker than coeval ones
446 due to their structural heritage and thermal history, and are assumed to be ‘attenuated’ (*sensu*
447 Fosdick, Graham & Hilley, 2014). The Farmington Terrane was formed as a mobile belt
448 between Archean GCB and WT and underwent at least one event of intense thermal
449 metamorphism during Mesoproterozoic time (Mueller *et al.* 2011; Lund *et al.* 2015) Younger
450 occurrences of similar events until Early Triassic time cannot be ruled out, especially given
451 the Bouguer gravity anomaly hints of underplating dense material (see Section 4.b). The
452 Farmington Terrane is therefore considered here as a ‘thermally attenuated weak’ lithosphere
453 (Fig. 11).

454 Due to the lithospheric heterogeneity of the basement, the role of the boundary
455 lithospheric faults can be considered as essential. Neoproterozoic–Palaeoproterozoic terranes are
456 limited by mega-shear zones along with deep (nearly) vertical crustal and/or lithospheric
457 faults (Figs 10d, 11). Terranes in the SFB display some characteristics (e.g. dimension,
458 geometry) that are similar to the terranes associated with the Neoproterozoic–Palaeoproterozoic
459 accretionary orogens (e.g. Chardon, Gapais & Cagnard, 2009, fig. 2; Cagnard, Barbey &
460 Gapais, 2011, fig. 9). These lithospheric and crustal accidents have therefore been reactivated
461 since their Precambrian onset (e.g. Bryant & Nichols, 1988; Paulsen & Marshak, 1999).
462 Additionally, several authors (e.g. Eardley, 1939; Peterson, 1977) identified the presence of a
463 topographic basement highland (pale blue area in Fig. 12a, in colour online) near the junction
464 between the MT and the GCB/FT/WT during Palaeozoic time, separating the northern and
465 southern areas of marked sedimentary accumulation. Eardley (1939) first introduced this
466 feature as the ‘Northern Utah Highland’. Peterson (1977) highlighted its presence on his
467 palinspastic maps for the Palaeozoic stratigraphic record. Finally, this sedimentary and
468 topographic pattern seems to have been the same in this basin since Proterozoic time (Paulsen
469 & Marshak, 1999; Fig. 12a).

470 By the time of the initiation of the Sonoma orogeny, this difference in sedimentary
471 accumulation was well marked in Palaeozoic series (Peterson, 1977). For instance, about
472 6<5>km of marine sediments accumulated in the Permian Oquirrh Basin in the northern part
473 of the SFB (Fig. 12a; Yonkee & Weil, 2015), whereas the southern part of the SFB saw the
474 deposition of only several hundred metres of marine and terrigenous sediments (e.g.
475 c. 640<5>m in southwestern Utah; Rowley *et al.* 2005) during the same interval. The thick
476 Palaeozoic sedimentary series in northern and southern parts of the foreland (Peterson, 1977)
477 would have allowed the thrust belt to propagate, while the presence of the topographic
478 basement highland characterized by a reduced sedimentary cover should have triggered the
479 formation of a lateral ramp and a recess in the central part of the front (Fig. 12a). The
480 presence of the topographic high is attested by the occurrence of shallow conglomerates in
481 the western part of the SFB within the PTU-Smithian interval (Fig. 12a, b; e.g. Gabrielse,
482 Snyder & Stewart, 1983; Lucas & Orchard, 2007; Jattiot *et al.* in press). Previous
483 reconstruction of the GA thrust front also accounted for the presence of a recess in the central
484 part of the thrust front (e.g. Dickinson, 2006, 2013). Moreover, this mechanism underlying
485 the observed differential propagation has been proposed by Paulsen & Marshak (1999) for
486 the Sevier thrust-and-fold belt which shows the presence of a lateral ramp in its central part
487 (Fig. 2). This was explained by the pre-deformational sedimentary thicknesses pattern
488 showing thrusts propagating further when emplaced upon a thicker sedimentary cover
489 (Figs 2, 12a; Paulsen & Marshak, 1999, fig. 7). It is worth noting that both the lateral ramps
490 of the Sevier and Golconda thrust-and-fold belt are located close to and along the lithospheric
491 boundary between the MT and FT/WT (Figs 2, 12a).

492 The GA heterogeneity may therefore have played a role, complementary to the
493 basement heritage, over the flexural response of the SFB. However, due to the scarcity of
494 allochthon remnants, a numerical model is required to decipher its potential role.

495 4.d. Simulating the flexural response of the basin

496 All the data discussed above have been integrated in a 2D numerical flexural model.
497 This approach allows us to quantify in a predictive way the flexural behaviour of the basin in
498 relation to its basement heritage.

499 4.d.1. Numerical approach and setup

500 The 2D plane stress flexural models have been solved with a finite element method
 501 code written in Matlab® (Le Pourhiet & Saleeby, 2013; Moreau *et al.* 2015). It solves

$$502 \quad \nabla^2 (D \nabla^2 \omega) = g (\rho_m - \rho_i) + q \quad (1)$$

503 for flexural deflection ω of a thick elastic plate (Reissner–Mindlin approximation) using
 504 bilinear isoparametric elements with under integration technique for the shear terms
 505 (Zienkiewicz & Taylor, 2005). In Equation (1) the rigidity of the plate D , defined

$$506 \quad D = \frac{ET_e^3}{12(1-\nu^2)},$$

507 depends solely on the effective elastic thickness T_e as the plate Young’s modulus E and
 508 Poisson’s ratio ν are fixed at 80<5>GPa and 0.25, respectively (Burov & Diament, 1995).
 509 The topographic loads $q = \rho g h$ account for the thickening h resulting from the orogeny and
 510 are computed using a density $\rho = 2700<5>\text{kg m}^{-3}$. The mantle restoring forces are computed
 511 assuming a density $\rho_m = 3300<5>\text{kg m}^{-3}$, while the infill is considered to be sediments of
 512 density $\rho_i = 1600<5>\text{kg m}^{-3}$. We arbitrarily attributed a constant height $h = 1500<5>\text{m}$ to the
 513 topographic load as we concentrate on the effect of heterogeneities of the allochthon
 514 morphology and rheology of the basement only. These initial parameters are summarized in
 515 Table 2.

516 The models are 907<5>km wide in the x direction, chosen to be normal to the trend of
 517 the orogenic belt, and 1166<5>km in the y direction. We assume that isostatic compensation
 518 is achieved underneath the orogen and, accordingly, we set the curvature normal to the right
 519 side to zero, $\partial\omega/\partial x = 0$. As the orogen is very long compared to the region where flexural
 520 subsidence is analysed, we enforce cylindrical boundary conditions on the side of normal y
 521 ($\partial\omega/\partial y = 0$). On the right boundary, that is, far from the orogeny, the effect of topographic
 522 loading can be considered null, corresponding to $\omega = 0$.

523 In this model, we used $T_{e1} = 90<5>\text{km}$ for the ‘strong’ GCB, WT, MT and YT
 524 lithospheres (Table 2), which is a good approximation for cratonic T_e (Watts, 1992). The
 525 ‘weak-attenuated’ FT is expected to show a contrasted lower T_e value due to its assumed

526 rheological weaknesses. This value was set at $T_{e2} = 30$ km (Table 2; e.g. Leever *et al.*
527 2006).

528 4.d.2. Model results

529 Figure 13 shows that the southern part of the front is reconstructed as less propagated
530 into the foreland than the northern part (Fig. 12a; see Dickinson, 2006, 2013). In this model,
531 the lateral ramp is spatially restricted along the limit between the FT/WT and MT (Fig. 13a).
532 The northern part, emplaced mainly above the ‘weak’ FT and in front of the largest part of
533 the GA, presents a narrower foredeep with $\lambda \approx 250$ km (Fig. 13a, b). The steep foredeep is
534 bordered by a well-expressed forebulge emplaced close to the FT/WT boundary (Fig. 13a;
535 *XX'* in Fig. 13b). The southern part of the foreland is set upon ‘strong’ lithospheres (MT and
536 YT) in front of the smallest and recessed parts of the GA (Fig. 13a, c). The foredeep in this
537 part of the model is larger, with $\lambda \approx 320$ km, and its profile (*YY'* in Fig. 13c) also exhibits
538 a weaker topography than in the northern part. We also notice the presence of a barely
539 expressed forebulge in this area (Fig. 13a, c).

540 The dichotomy between the northern and southern parts is especially obvious on a N–S
541 transect (*ZZ'* in Fig. 13d). A shallow southern sub-basin with a gentle northwards dip (<
542 *c.* 250 m deep) is identified, as well as a northern deeper basin with steep borders
543 (*c.* 600 m deep). The limit between the northern and southern parts appears relatively
544 close to the MT/FT boundary (Fig. 13d), suggesting a significant role for lithospheric
545 boundaries in the differential flexuration of the SFB. This N–S differentiation is found not
546 only in the foreland, but also within the allochthon itself as its simulated elevation is not
547 continuous along its front (Fig. 13a). Two areas of important elevations (>1200 m) can be
548 observed on both the northern and southern sides of the GA recess. This positive relief could
549 have contributed as a significant source of terrigenous material, then being deposited in the
550 proximal foreland.

551 5. Discussion

552 Our results highlight the spatial differences in subsidence within the SFB, especially
553 between its northern and southern parts (Figs 8, 9). This differential subsidence is underlined
554 by variations in the sedimentary record (Figs 4, 5). In addition, a highland was probably

555 present in the central SFB and could physically have partly separated these two parts of the
556 basin.

557 5.a. Evidence for a foreland basin

558 The convex ‘lozenge shape’ (*sensu* Miall, 2010) of the isopach map (Fig. 8) and the
559 westwards-thickening pattern of the sedimentary record are in agreement with the common
560 asymmetric geometry of foreland basins (Fig. 8; DeCelles & Giles, 1996; Miall, 2010).
561 Additionally, the observed high-rate subsidence values (*c.* 100–500 m Ma^{-1}) agree with
562 foreland basin dynamics, even if these values are greater in magnitude than values generally
563 given in the literature for similar contexts (e.g. Xie & Heller, 2009). This difference in
564 magnitude is interpreted by considering that estimations from backstripping analyses are
565 generally proposed for continuous sedimentary series spanning several millions years, if not
566 several tenth of millions years (e.g. Xie & Heller, 2009). Over such long time intervals, the
567 subsidence rate values are less accurate. The high resolution of the timeframe used for the
568 SFB mirrors short-acting structural events in the basin. Similar ‘higher than average’ values
569 for subsidence rates have been calculated by Chevalier *et al.* (2003) and Lachkar *et al.* (2009)
570 using high-resolution biostratigraphic time-calibrations, and also by Roddaz *et al.* (2010)
571 with similar magnitude for the Miocene Amazonian Foreland Basin (*c.* 200–700 m Ma^{-1} ;
572 Roddaz *et al.* 2010). Moreover, values observed in the SFB (0.05–0.65 mm a^{-1}) are consistent
573 with yearly deposition rates indicated by Allen & Allen (2005) for foreland basins (0.2–
574 0.5 mm a^{-1}). Finally, the convex-up shape of the tectonic subsidence curves (Fig. 9f) is
575 diagnostic of foreland basins and corresponds to the progressive flexural response of the
576 lithosphere to the topographic load and/or sedimentary infill of the basin overtime (Angevine,
577 Heller & Paola, 1990; Allen & Allen, 2005; Xie & Heller, 2009).

578 In the SFB, the topographic load is exerted by the GA. This allochthon has been
579 emplaced on the North American continental margin, as evidenced by the geochemical
580 signature of the Koipato Formation volcanics (Early Triassic) originating from the partial
581 melting of a Palaeoproterozoic continental crust (likely the Mojave Terrane; Vetz, 2011).

582 The observed spatial heterogeneity of the sedimentary thickness in the SFB (Figs 4, 8)
583 and the much higher tectonic subsidence rate detected in the northern part of the basin
584 (*c.* 500 m Ma^{-1} v. *c.* 100 m Ma^{-1} in the southern part; Fig. 9f) are striking and raise the

585 question of the controlling factor(s) responsible for this phenomenon, especially for such a
586 short interval (*c.* 1.3–5 Ma).

587 5.b. Potential underlying mechanisms for observed variations in flexural 588 subsidence

589 Spatial variations in subsidence within the SFB may result from different mechanisms
590 that are inherent to the flexural nature of the foreland basin: (1) the sedimentary overload
591 provoked by the continuous filling of the basin over time; (2) the spatial heterogeneity of the
592 GA (topography and shape of the load); and/or (3) the differential flexural response of the
593 lithosphere to this topographic load and linked to the rheology of the basement.

594 Considering point (1) above, in some cases the distributed vertical load exerted by the
595 sedimentary filling of the basin might affect and amplify the flexuration in foreland basins
596 over time (Shanmugam & Walker, 1980; Beaumont, 1981; Cardozo & Jordan, 2001; Allen &
597 Allen, 2005). As this load depends mainly on the sedimentary fluxes and density of the
598 filling, a denser deposited material leads to a more important flexuration of the lithosphere, as
599 modelled by Angevine, Heller & Paola (1990) and Fosdick, Graham & Hilley (2014). The
600 southern part of the SFB, characterized by low subsidence rates, exhibits coarse clastic
601 sedimentation in the Moenkopi Group with the presence of conglomerates and sandstones
602 (Figs 3a, 4, 5b, c, e, 12; e.g. Gabrielse, Snyder & Stewart, 1983; Olivier *et al.* 2016) of
603 density 2.5–2.8 kg cm⁻³ (Manger, 1963; McCulloh, 1967; Sclater & Christie, 1980;
604 Tenzer *et al.* 2011). The top of the Moenkopi Group consists of thick microbial limestone
605 beds (Figs 3a, 4, 5e; e.g. Olivier *et al.* 2014, 2016; Vennin *et al.* 2015). These limestones bear
606 a density of *c.* 2.6–2.8 kg cm⁻³ (Manger, 1963; McCulloh, 1967; Sclater & Christie, 1980;
607 Tenzer *et al.* 2011). In contrast, the northern part which is characterized by high subsidence
608 rates, is dominated by marine siltstones of the Dinwoody and Woodside Formation (Figs 3a,
609 4, 5g; e.g. Kummel, 1954, 1957). The density of this type of sediment is of 2.3–
610 2.7 kg cm⁻³ (Manger, 1963; Sclater & Christie, 1980; Tenzer *et al.* 2011). Based on these
611 data, the sedimentary filling should have had a higher impact on the flexuration in the
612 southern part of the basin. However, we show that the most important subsidence during the
613 PTU-Smithian interval took place in the northern part of the SFB (Figs 8, 9). Moreover, the
614 difference between tectonic and total subsidence mainly consist of the local isostasy and

615 compaction of the sediments (Allen & Allen, 2005). With the tectonic subsidence being the
616 most important component of the total subsidence in the SFB (Fig. 9a), this argues for a weak
617 potential role of the sedimentary load. The sedimentary overload therefore cannot be a major
618 controlling factor explaining the differential flexuration observed within the basin.

619 Regarding points (2) and (3) above, while it is possible to discuss the role of the
620 sedimentary overload using only field-based data, interpretations of the allochthon
621 heterogeneity and the basement rheological behaviour require an additional model approach.
622 We combine these in the following discussion. To that purpose, we used three different
623 scenarios (Fig. 14) with the same initial setup (Section 4.d; Table 2) except for the x and y
624 dimensions of the model that are set to 2000 km in the x direction and 1000 km in the
625 y direction to avoid border effects.

626 The first scenario tests the impact of a rheologically heterogeneous basement loaded by
627 a homogeneous allochthon (Fig. 14a). The rigidity of the terrane controls its capacity to
628 flexure. The shape of ensuing flexural foreland basins and the distribution of their
629 sedimentary records are therefore a direct consequence of the rheological behaviour of the
630 basement (Angevine, Heller & Paola, 1990; Watts, 1992; Cardozo & Jordan, 2001; Allen &
631 Allen, 2005; Leever *et al.* 2006; Fosdick, Graham & Hilley, 2014). Upon the high-rigidity
632 part of the basement (T_{e1}), a wide foreland ($\lambda_1 \approx 250$ km) develops with a well-expressed
633 convex shape in map view and a barely expressed forebulge. Upon the low-rigidity parts of
634 the basement (T_{e2}), a narrower foreland ($\lambda_2 \approx 110$ km) is structured with a more
635 pronounced forebulge. This is in agreement with the SFB observations. However, a N–S
636 transect (aa' , Fig. 14a) shows that the wider area of the foreland basin is deeper than
637 observed in the field and that only one high-relief area is individualized within the central
638 part of the allochthon. Even if the rigidity does play a role in the development of the flexural
639 foreland basin, as commonly assumed in the literature (Angevine, Heller & Paola, 1990;
640 DeCelles & Giles, 1996; Cardozo & Jordan, 2001; Allen & Allen, 2005; Leever *et al.* 2006;
641 Miall, 2010; Fosdick, Graham & Hilley, 2014), our results indicate that a rheological
642 difference is not enough to control the variations in SFB.

643 The second scenario uses a heterogeneous topographic load exerted by the allochthon
644 upon a homogeneous ‘strong’ lithosphere ($T_e = 90$ km; Fig. 14b). The heterogeneity in the

645 allochthon is introduced in the form of a *c.* 100<5>km wide recess (i.e. a lateral ramp) along
646 its front. The foreland basin shows a larger area ($\lambda_1 \approx 180<5>$ km) in front of the lateral ramp
647 compared to the northern and southern parts ($\lambda_2 \approx 100<5>$ km). Moreover, a N–S transect (*bb'*
648 in Fig. 14b) shows that the narrow northern part of the basin is deeper than in front of the
649 recess. An important relief is also formed in the corners of the allochthon on both lateral
650 borders of the recess. This is in agreement with SFB observations. However, the overall
651 shape of the foreland basin is rather concave and enters in the recess significantly. Even if the
652 morphology of the allochthon plays a role in the development of the foreland basin, this
653 numerical scenario shows marked differences with the SFB.

654 The third scenario combines both previously tested heterogeneities (Fig. 14c). The
655 graphic output exhibits a wider foreland ($\lambda_1 \approx 350<5>$ km) emplaced above the ‘strong’
656 lithosphere in front of the recess, and a narrow foreland ($\lambda_2 \approx 100<5>$ km) above ‘weak’
657 lithospheres. This model reproduces well the convex shape of the foreland basin with a
658 marked forebulge development upon ‘weak’ lithospheres, whereas it is less pronounced upon
659 the strong lithosphere. Moreover, a N–S transect (*cc'* in Fig. 14c) highlights a deeper area
660 upon the ‘weak’ lithosphere. Finally, a prominent relief of the allochthon is observed on both
661 corners bordering the recess.

662 To summarize, from the three possible mechanisms proposed to explain the origin of
663 the differential flexural subsidence in the SFB, only the combined effect of the heterogeneous
664 rheology of the basement and the spatial heterogeneity of the GA can be considered as the
665 major controlling factors.

666 5.c. Combined outcomes of heterogeneities over differential subsidence

667 Our field data highlight the contrasted subsidence between the northern and southern
668 parts of the SFB. The numerical model provides a complement to discuss the potential
669 combined outcomes of rheology and allochthon heterogeneities. Congruent features between
670 the numerical model of the SFB (Fig. 13), the tested scenarios (Fig. 14) and field data
671 (Fig. 15) indeed argue for a major controlling role of the allochthon spatial heterogeneities
672 and of the basement rheological behaviour on the formation and development of the SFB
673 during Early Triassic time. As these two parameters are directly linked to the age, nature and

674 pattern of the basement terranes, the lithosphere heritage likely controls the flexuration and
675 therefore the subsidence variations documented for the Early Triassic SFB.

676 Combining all field data and numerical simulations, a model of the SFB is proposed in
677 Figure 15. The northern part of the basin (section *AA'*) is characterized by a narrow foredeep
678 ($\lambda \approx 250$ km) with a high-rate tectonic subsidence ($c. 500$ m Ma⁻¹) and high
679 sedimentary thickness (up to $c. 550$ m of mostly fine siltstones deposits), which is located
680 upon the 'weak/attenuated' Farmington Terrane and in front of the largest reconstructed part
681 of the GA. The postulated wedge-top and forebulge are located above the 'strong' Archean
682 lithospheres, that is, the GCB and WT, respectively. The southern part of the SFB exhibits a
683 large foredeep ($\lambda \approx 500$ km, section *BB'*) with a relatively low-rate tectonic subsidence
684 ($c. 100$ m Ma⁻¹) and a reduced sedimentary thickness (up to $c. 250$ m of mixed
685 limestones and coarse clastic deposits). This part of the SFB is emplaced upon the 'strong'
686 lithospheres of the Palaeoproterozoic MT and YT, in front of the thinnest reconstructed part
687 of the GA. The southern SFB also shows a reduced postulated wedge-top to the west and a
688 barely expressed forebulge to the east. These spatial variations in flexural subsidence and
689 their good agreement with limits of the terranes composing the SFB basement are also
690 evident along a N–S transect (section *CC'*). The spatial separation between the shallow and
691 gently dipping southern part of the SFB and the deep and steep northern part is obvious. This
692 separation is located close to the boundary between MT and FT.

693 6. Conclusion

694 In this study, we used an integrated approach to decipher the major role of the
695 lithospheric heritage over the differential sedimentary deposition in the Sonoma Foreland
696 Basin during Early Triassic time. Our approach used both field-based sedimentary data,
697 calibrated within a highly resolved biostratigraphic framework, and numerical model to test
698 the influence of several potential controlling factors. Palinspastic reconstructions were also
699 performed to obtain an accurate palaeogeographic context.

700 Using high-resolution temporal data, the subsidence analyses help to identify the main
701 controlling factors at the origin of the spatial variations of the Early Triassic sedimentary
702 record in the SFB. The sedimentary overload cannot satisfactorily explain the observed

703 variations in thickness of the sedimentary record throughout the basin. The combined effects
704 of the contrasted lithospheric strength of the terranes ('weak' v. 'strong' lithospheres)
705 composing the basement of the basin, and the spatial heterogeneity of the Golconda
706 Allochthon (with the presence of a lateral ramp within the belt), best explain a differential
707 flexural response of the SFB basement to the emplacement of the allochthon. Such a
708 differential flexural response ultimately controls the overall geometry of the basin through
709 spatially heterogeneous tectonic subsidence rates: *c.* 100–500 m Ma⁻¹ in a wide southern part
710 upon a 'strong' lithosphere loaded by a recessed and thin (in map-view) front belt, v.
711 *c.* 500–1000 m Ma⁻¹ in a narrower northern part upon a 'weak/attenuated' lithosphere loaded by
712 a larger front belt. Although field data highlight the potential role of the rheological
713 behaviour of the basement based on observed differential subsidence rates, the numerical
714 model approach suggests a combined effect of the latter and of the spatial heterogeneity of
715 the allochthon.

716 As heterogeneities of the basement and in the morphology of the allochthon result from
717 the nature and history of the different lithospheric terranes that compose the basement, the
718 lithosphere heritage likely played a prime role in controlling the development of the Sonoma
719 Foreland Basin during Early Triassic time, and consequently generated the observed
720 variations of the sedimentary record through differential subsidence.

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726 Supplementary material

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1063 Figure 1. (Colour online) (a) Early Triassic location of the Sonoma Foreland Basin (SFB;
1064 after Brayard *et al.* 2013). (b) Simplified chronostratigraphy of the succession of structuring
1065 events in the studied area since Palaeoproterozoic time (after Oldow *et al.* 1989; Whitmeyer
1066 & Karlstrom, 2007; Dickinson, 2013). (c) Simplified map of the study area with location of
1067 the main structural elements discussed and mentioned in this work (after Bond *et al.* 1985;
1068 Walker, 1985; Dickinson, 2004, 2006, 2013; Vetz, 2011; Yonkee & Weil, 2015).

1069 Figure 2. (Colour online) Topographic map of the central part of current-day Sevier thrust-
1070 and-fold belt with accentuation of the Wyoming and Central Utah salients thrusts. A lateral
1071 ramp is present between the two salients (after Paulsen & Marshak, 1999).

1072 Figure 3. (Colour online) (a) Simplified litho- and chronostratigraphic subdivisions of the
1073 Early Triassic Sonoma Foreland Basin (SFB). This study encompasses the PTU-Smithian
1074 interval, with Spathian complement for the subsidence analysis. Main ammonoid markers
1075 used in this study are the *Anasibirites* beds and the *Columbites* beds. Radiometric ages: (1)
1076 from Burgess, Bowring & Shen (2014); (2) and (3) from Galfetti *et al.* (2007). (b) State map
1077 of the study area showing current location of the 43 studied sections, from both literature data
1078 (open circles) and field data (grey circles). Complete GPS coordinates and references are
1079 given in online Supplementary Table S2. Red outlines highlight the sections used for the
1080 subsidence analysis, and selected for their completeness, temporal resolution and spatial
1081 distribution. Sections detailed in Figure 4: SC: Sheep Creek; HS: Hot Springs; LWC: Lower
1082 Weber Canyon; CR: Confusion Range; T: Torrey area; PR: Pahvant Range; M: Minersville;
1083 RC: Rock Canyon.

1084 Figure 4. Biostratigraphic correlation based on the *Anasibirites* and *Columbites* beds
1085 observed in 8 of the 43 studied sections, illustrating the discrepancy in sedimentary thickness

1086 between the southern and northern parts of the Sonoma Foreland Basin (with simplified
1087 lithology). Base of the sections corresponds to the regionally recognized Permian–Triassic
1088 unconformity (Brayard *et al.* 2013).

1089 Figure 5. (Colour online) Photographs of different outcrops in the SFB, showing variations in
1090 dominant lithologies and sedimentary thicknesses encountered throughout the basin. (a)
1091 Panorama of Rock Canyon (RC) outcrop, showing the plurimetric beds of conglomerates
1092 from the basal Moenkopi Group. (b) Detail photograph of the conglomerate from Rock
1093 Canyon. (c) Photograph of the terrigenous red beds of the Moenkopi Group at Lower Weber
1094 Canyon (LWC). (d) Panorama of the limestones beds of the Thaynes Group limestones at
1095 Lower Weber Canyon. (e) Panorama of the Moenkopi Group at Minersville (M), showing
1096 succession of terrigenous red beds and microbial limestones. (f) Panorama of the transition
1097 between Moenkopi and Thaynes Group showing succession of microbial limestones and
1098 bioclastic limestones at Minersville. (g) Photograph of the marine siltstones of the Dinwoody
1099 and Woodside Formation at Hot Springs (HS). (h) Panorama of the Hot Springs section,
1100 showing succession of limestone levels of the Thaynes Group bioclastic limestones.

1101 Figure 6. (Colour online) Present-day and retrodeformed (for the PTU-Smithian interval)
1102 configurations for two regional cross-sections in the (a) northern and (b) southern parts of the
1103 Sonoma Foreland Basin, illustrating the method used for palinspastic reconstruction (after
1104 Groshong, 2006). Balanced cross-sections adapted from (a) Yonkee & Weil (2010) and (b)
1105 DeCelles & Coogan (2006) illustrate the retrodeformation process used to estimate the value
1106 of the tectonic transport, and therefore the approximate original location of the sections
1107 during the studied interval. Triassic series (highlighted layers) are used as the basis for the
1108 retrodeformation process and are horizontalized between the designated Pin and Loose lines
1109 (see text for details). The two cross-sections are located in Figure 7.

1110 Figure 7. (Colour online) Map representing the present-day location of the studied sections
1111 (dots) and their reconstructed position (open circles) obtained after retrodeformation.
1112 Positions of balanced cross-sections (a) and (b) illustrated in Figure 6 are also indicated. The
1113 present-day Sevier Thrust-and-Fold Belt (TFB; after Yonkee *et al.* 2014) is the main
1114 structural element responsible for tectonic transport during post-Triassic times. Black arrows
1115 represent the retrodeformation values applied from the present-day location of the studied
1116 sections. Seven sectors of similar estimated tectonic transport are delimited by dashed lines
1117 (see Table 1). Sector 1: Sevier foreland; Sector 2: Wyoming salient, northern part; Sector 3:
1118 Wyoming salient, central part; Sector 4: Wyoming salient, southern part; Sector 5: Central
1119 Utah salient, northern part; Sector 6: Central Utah salient, southern part; Sector 7: Sevier
1120 hinterland.

1121 Figure 8. (Colour online) Isopach map of the sedimentary thicknesses recorded for the PTU-
1122 Smithian interval, showing marked differences in sedimentary thicknesses between northern

1123 and southern Sonoma Foreland Basin. The studied sections are shown at their palaeolocation
1124 (Fig. 7). The reconstructed Golconda Allochthon Thrust Front during the PTU-Smithian
1125 studied interval is also indicated (modified from Dickinson, 2013; see also Fig. 12). The
1126 position of the wedge-top is based on variations in the sedimentary thicknesses and on
1127 geophysical data (Fig. 10).

1128 Figure 9. (Colour online) Subsidence analysis results obtained for the PTU-Smithian interval
1129 and early Spathian time using 1D backstripping (Steckler & Watts, 1978; Van Hinte, 1978;
1130 Allen & Allen, 2005). Locations of sections are given in Figure 3b. Ages for the bottom and
1131 top boundaries of the Smithian are interpolated from ammonoid biozone durations (after
1132 Brühwiler *et al.* 2010). Sea-level curve after Haq, Hardenbol & Vail (1988). *Ana.*:
1133 *Anasibirites* beds; *Col.*: *Columbites* beds. Radiometric ages from (1) Burgess, Bowring &
1134 Shen (2014); (2) and (3) Galfetti *et al.* (2007). Subsidence analysis for: (a) Confusion Range
1135 (CR) section; (b) Pahvant Range (PR) section; (c) Sheep Creek (SC) section; (d) Hot Springs
1136 (HS) section. (e) Total subsidence curves for all the CR, PR, SC and HS sections and
1137 associated dominant lithologies are indicated for each subinterval. (f) Tectonic subsidence
1138 curves for the CR, PR, SC and HS sections and associated mean tectonic subsidence rates. (e)
1139 and (f) allow two distinct subsidence dynamics to be discriminated between the southern and
1140 northern parts of the SFB.

1141 Figure 10. (Colour online) (a) Bouguer gravity anomaly map of the Sonoma Foreland Basin
1142 and its surroundings (in mGal; after Kucks, 1999). Notable moderate gravity anomalies are
1143 highlighted by a white contour. SRP: Snake River Plain; FA: Farmington Anomaly; SA:
1144 Southern Anomaly. Black lines represent the interpreted remnants of the main geophysical
1145 accidents, and limits between crustal features. (b) Aeromagnetic anomaly map of the Sonoma
1146 Foreland Basin and its surroundings (in nT; after Bankey *et al.* 2002). Black lines highlight
1147 areas of contrasted magnetic signatures: SRP: Snake River Plain; SZ: Southern magnetic
1148 Zone; CZ: Central magnetic Zone; NEZ: North-Eastern magnetic Zone; NZ: Northern
1149 magnetic Zone. (c) Map of the spatial location of the radiochronological ages (U/Pb ages)
1150 after: (1) Foster *et al.* 2006; (2) Fan *et al.* 2011; (3) Mueller *et al.* 2011; (4) Nelson, Hart &
1151 Frost, 2011; (5) Strickland, Miller & Wooden, 2011). Superimposed red dots indicate
1152 Mesoproterozoic metamorphism episodes (Mueller *et al.* 2011). (d) Map of basement terranes
1153 of the SFB according to their age and nature, with Archean terranes (pale blue),
1154 Palaeoproterozoic terranes (pale green) and Mesoproterozoic mobile belt (pale red). FT:
1155 Farmington Terrane; GCB: Grouse Creek Block; MT: Mojave Terrane; WT: Wyoming
1156 Terrane; YT: Yavapai Terrane.

1157 Figure 11. (Colour online) Map of the SFB basement (cf. Fig. 10d) after their heritage and
1158 therefore their rheological behaviour. Archean Grouse Creek Block and Wyoming Terrane,
1159 Palaeoproterozoic Mojave Terrane and Yavapai Terrane are considered 'strong' lithospheres
1160 with an important rigidity (pale blue), while the Mesoproterozoic mobile belt Farmington

1161 Terrane is considered a ‘thermally attenuated weak’ lithosphere due to its lesser rigidity (pale
1162 red).

1163 Figure 12. (Colour online) (a) Simplified map showing the position of the Uinta recess
1164 (lateral ramp) and Wyoming and Central Utah salients (frontal ramps) of the present-day
1165 Sevier TFB (after Paulsen & Marshak, 1999; Yonkee & Weil, 2010) and reconstructed
1166 Golconda Allochthon front and associated recess (lateral ramp). Sedimentary pattern since
1167 Proterozoic time shows two high accommodation zones separated by a topographic high
1168 close to the terrane boundaries (Peterson, 1977, Bryant & Nichols, 1988; Paulsen & Marshak,
1169 1999). Palaeolocation of Permian Oquirrh Basin (e.g. Yonkee & Weil, 2015) and documented
1170 PTU-Smithian conglomerates in the western SFB (e.g. Gabrielse, Snyder & Stewart, 1983;
1171 Lucas & Orchard, 2007) are also included on the map. Red lines indicate limits of the
1172 basement terranes (cf. Fig 9d). (b) Photograph (courtesy of Hugo Bucher, Zürich) of the
1173 conglomerates found in the area delimited in (a), presumably a product of western relief
1174 dismantlement.

1175 Figure 13. (Colour online) Numerical model of the SFB after the reconstructed
1176 palaeogeography and terranes map (cf. Figs 11, 12) with an heterogeneous basement (‘strong’
1177 v. ‘thermally attenuated weak’ lithospheres) and an heterogeneous allochthon (recessed area
1178 in central part of the front). (a) Simulated map of the SFB. Thin black lines indicate the
1179 position of the 2D profiles; red lines indicate limits of the basement terranes (cf. Fig 10d). (b)
1180 2D W–E profile of the northern part of the SFB model. The narrow foredeep is emplaced
1181 upon the ‘thermally attenuated weak’ FT and is bordered by a well expressed forebulge. (c)
1182 2D W–E profile of the southern part of the SFB model. The wider foredeep is emplaced upon
1183 the ‘strong’ MT, and is bordered by a barely expressed forebulge. (d) 2D N–S profile of the
1184 SFB model. The two northern and southern parts of the basin are individualized with a limit
1185 near the MT/FT boundary.

1186 Figure 14. (Colour online) Numerical models showing the effects of the heterogeneities of the
1187 basement and of the topographic load over the formation of a foreland basin. Dashed lines
1188 represent an area analogue to the SFB configuration. (a) Scenario using a heterogeneous
1189 basement with contrasted elastic thicknesses ($T_{e1} = 3 \times T_{e2}$) and a homogeneous allochthon. A
1190 large convex foreland is formed upon the most rigid lithosphere. (b) Scenario using a
1191 heterogeneous allochthon with a *c.* 100–500 km wide recess (lateral ramp) and a homogeneous
1192 fixed T_e lithosphere. A slightly wider concave foreland is formed within the recessed area and
1193 a cornering relief appears on both sides of the recessed area in the allochthon. (c) Scenario
1194 showing the combined effect of a heterogeneous basement with contrasted elastic thicknesses
1195 ($T_{e1} = 3 \times T_{e2}$) and a heterogeneous allochthon with a *c.* 100–500 km wide recess (lateral ramp).
1196 A much wider convex foreland is formed within the recessed area upon the rigid lithosphere,
1197 and a cornering relief on both sides of the recess in the allochthon is also visible.

1198 Figure 15. (Colour online) Cross-sections of the Sonoma Foreland Basin (SFB) illustrating
1199 variations in the subsidence and sedimentary accumulation pattern during the PTU-Smithian
1200 interval. The Golconda Allochthon (GA) is the main topographic load on the lithosphere
1201 (Dickinson, 2006, 2013; Marzolf, 1993); the postulated wedge-top is also represented. (*AA'*)
1202 W–E cross-section in the northern part of the basin exhibiting a narrow foreland with a high-
1203 rate tectonic subsidence with a developed silty and limestone sedimentation over the
1204 Mesoproterozoic ‘thermally attenuated weak’ Farmington Terrane (FT). (*BB'*) W–E cross-
1205 section in the southern part of the Sonoma Foreland Basin showing a wide foreland with a
1206 low-rate tectonic subsidence, forming a reduced deposition of mainly terrigenous clastic
1207 series upon the Palaeoproterozoic ‘strong’ Mojave Terrane (MT). A barely expressed
1208 forebulge borders this part of the SFB. (*CC'*). N–S cross-section of the basin, highlighting the
1209 differences between southern and northern parts of the SFB in terms of subsidence,
1210 sedimentation and geometry of the basin. The transition between these two parts is situated
1211 close to the terranes boundary between MT and FT. This area is postulated to be a basement
1212 topographic highland, as supported by the transition between southern terrigenous clastic
1213 series and northern silty sedimentation.

1214

1215 Table 1. Estimated tectonic transport values used for palinspastic reconstructions of each
 1216 sectors defined within the SFB, and associated references.

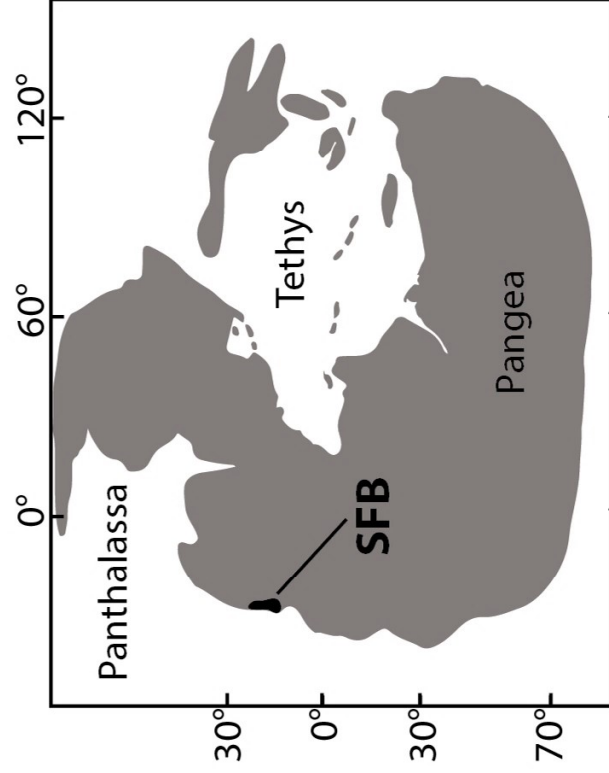
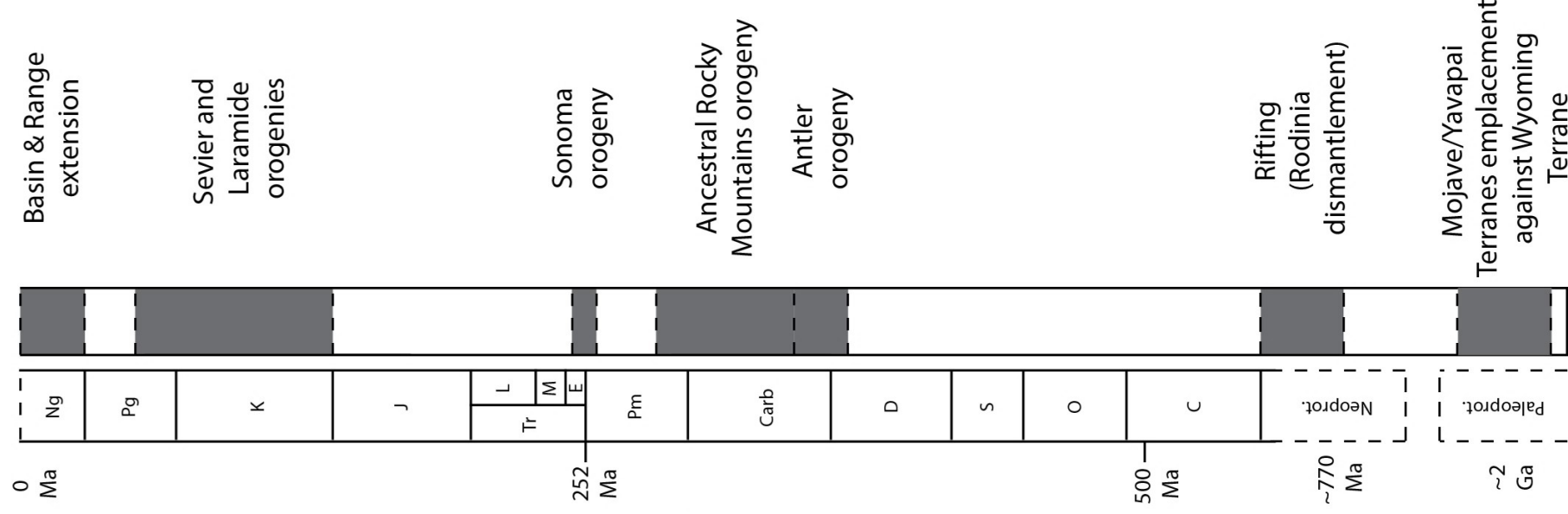
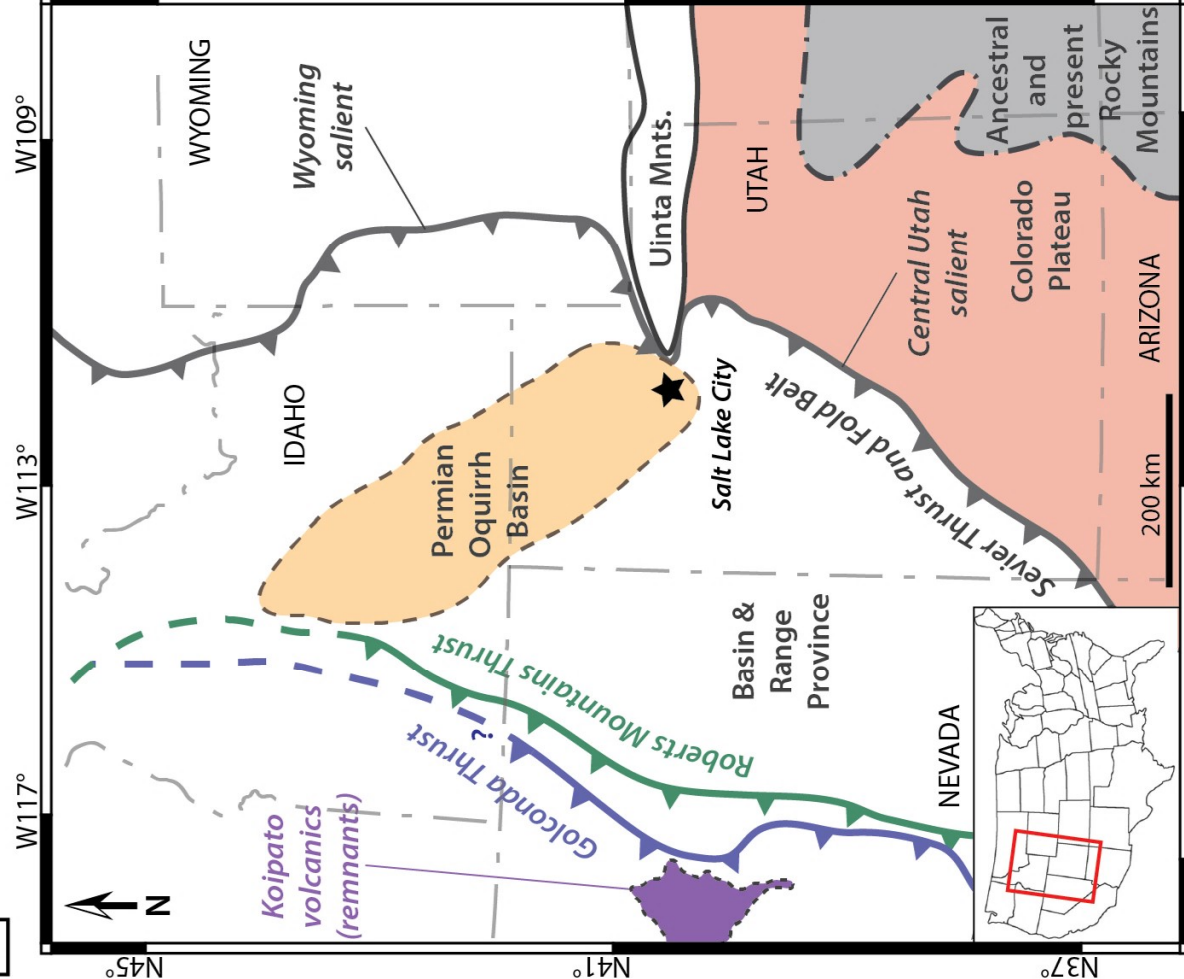
Sector	Estimated tectonic transport (km)	References
1 Sevier foreland	0	DeCelles & Coogan, 2006; Schelling <i>et al.</i> 2007; Yonkee & Weil, 2010; Yonkee <i>et al.</i> 2014
2 Wyoming salient, north part	<i>c.</i> 100	Paull & Paull, 1991; Yonkee & Weil, 2010
3 Wyoming salient, central part	140	Yonkee & Weil, 2010
4 Wyoming salient, south part	95	Yonkee & Weil, 2010
5 Central Utah salient, north part	100	Schelling <i>et al.</i> 2007
6 Central Utah salient, south sector	<i>c.</i> 75	DeCelles & Coogan, 2006; Schelling <i>et al.</i> 2007
7 Sevier hinterland, Basin & Range province	<i>c.</i> 80	Yonkee <i>et al.</i> 2014

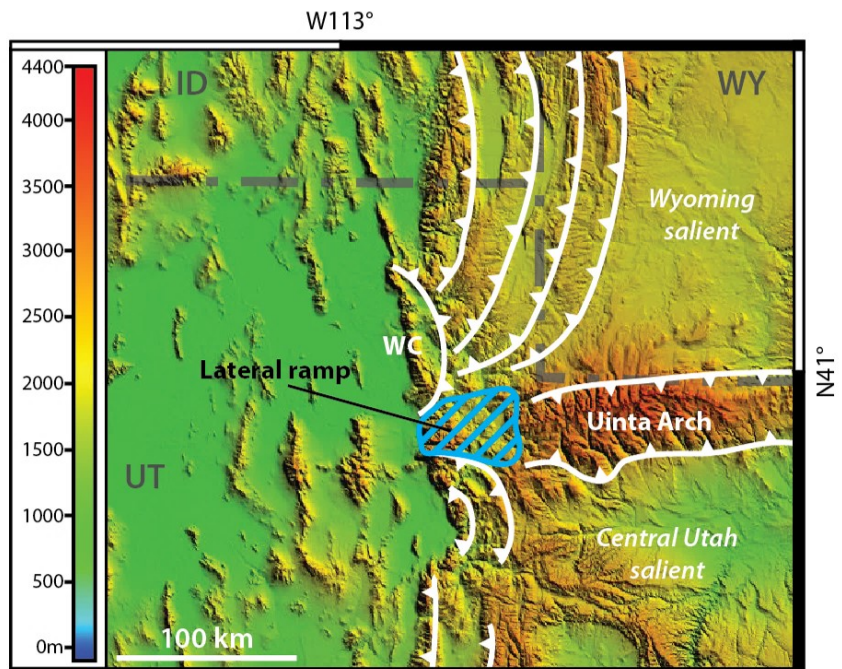
1217 Table 2. Summary of model parameters for the SFB and tested scenarii.

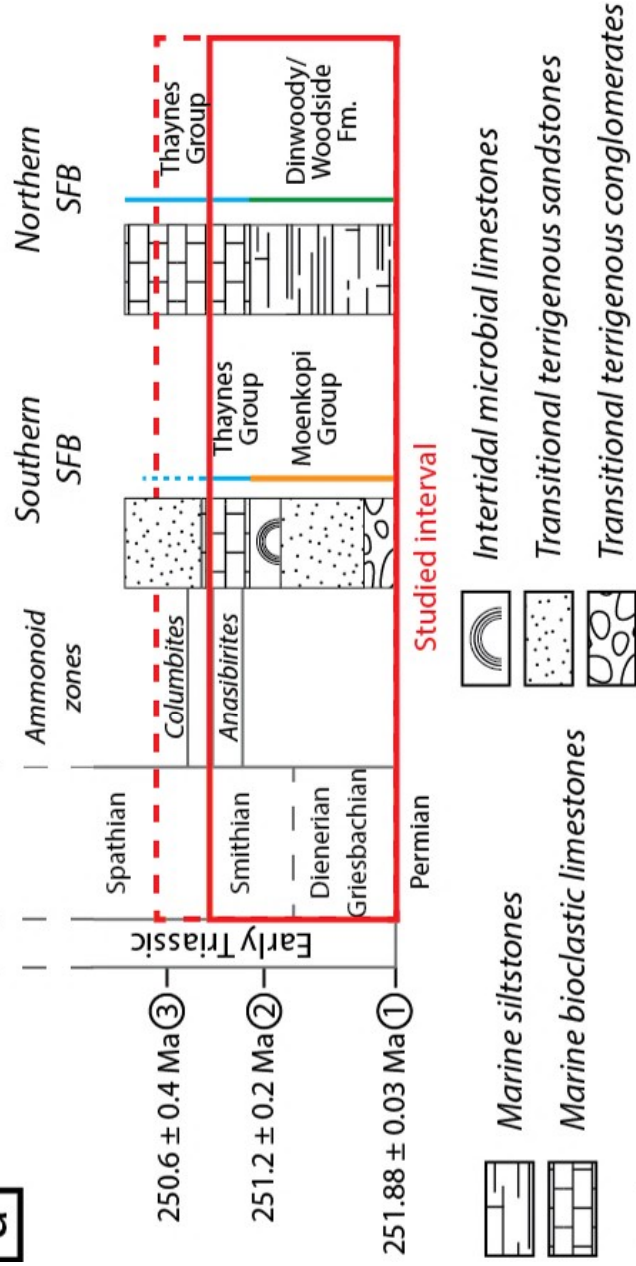
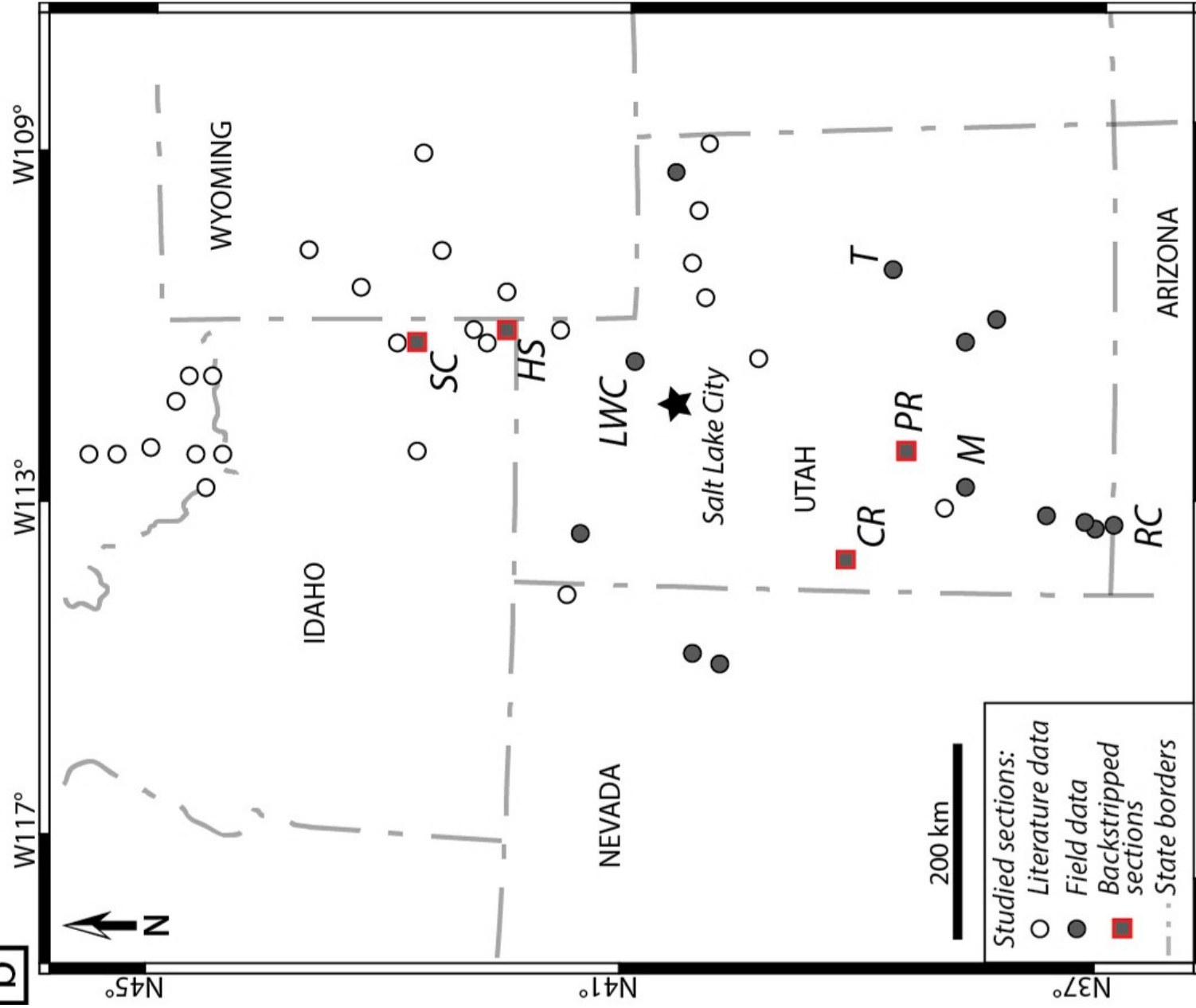
Parameter	SFB model (Fig. 13)	Heterogeneous basement scenario (Fig. 14a)	Heterogeneous allochthon scenario (Fig. 14b)	Combined heterogeneities (basement & allochthon; Fig. 14c)
Young's modulus E (GPa)	80	80	80	80
Poisson's ratio, ν	0.25	0.25	0.25	0.25
Elastic thickness of 'strong' lithosphere, T_{e1} (km)	90	90	90	90
Elastic thickness of 'weak' lithosphere, T_{e2} (km)	30	30	n/a	30
Loading parameters				
Allochthon thickening, h (m)	1500	1500	1500	1500
Density of topographic load, ρ_t (kg m^{-3})	2700	2700	2700	2700
Density of the mantle, ρ_m (kg m^{-3})	3300	3300	3300	3300
Density of the sedimentary infill, ρ_i (kg m^{-3})	1600	1600	1600	1600
Gravitational acceleration, g (m s^{-2})	9.81	9.81	9.81	9.81

1218

1219

a**b****Main structural events****c**



a**b**

Northern SFB

LWC HS SC

Southern SFB

RC

CR

PR

M

T

Col.

Col.

Ana.

Spathian

PTU-Smithian interval







Ana.

PTU-Smithian interval

100m

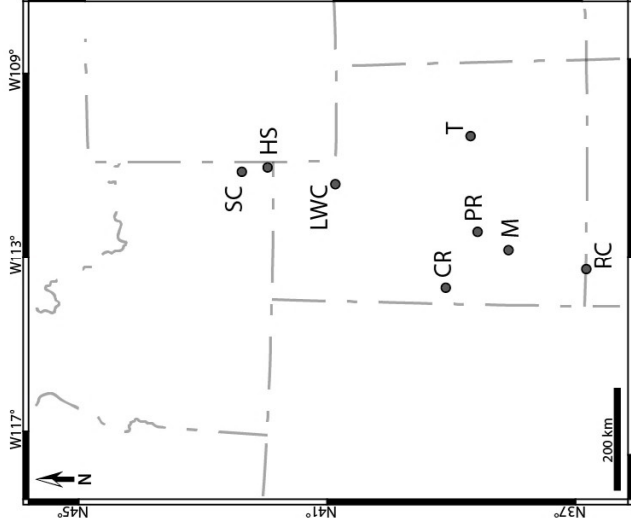
100m

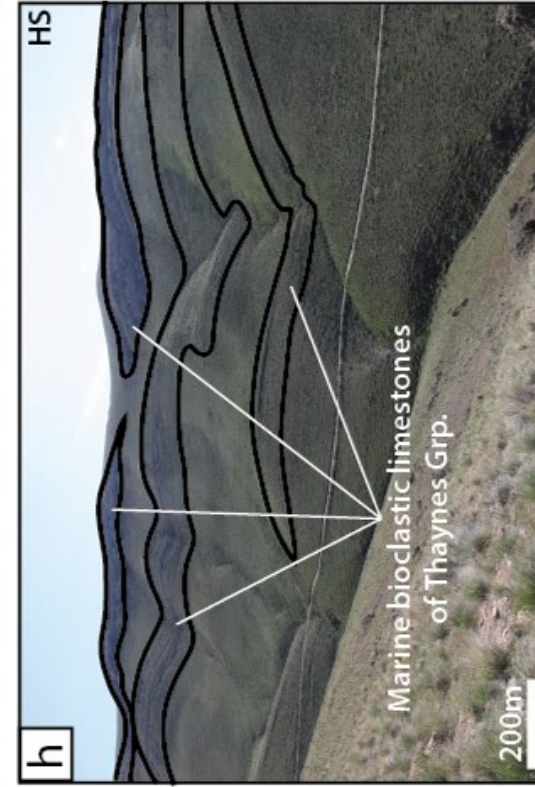
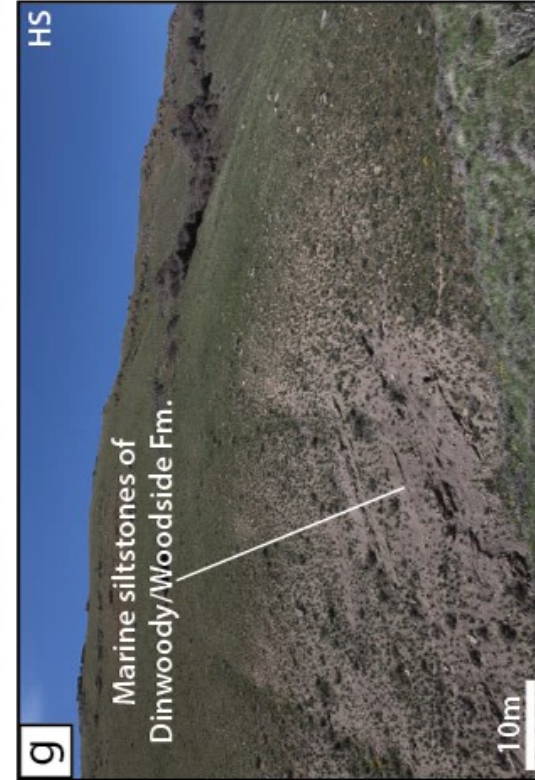
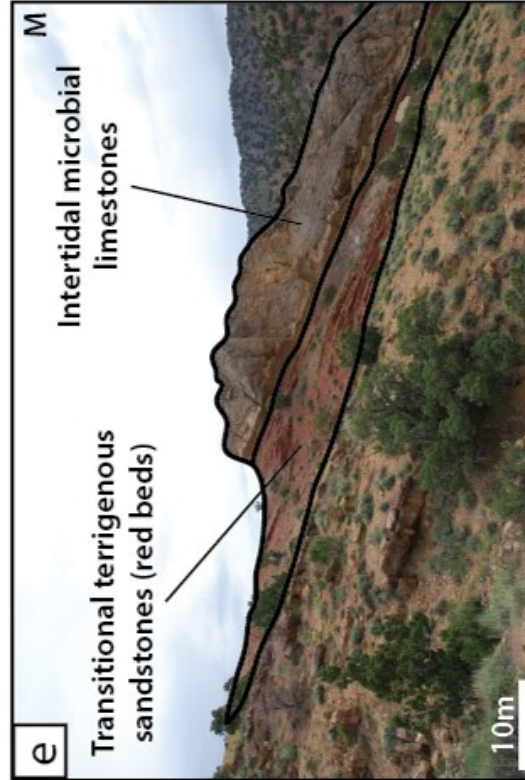
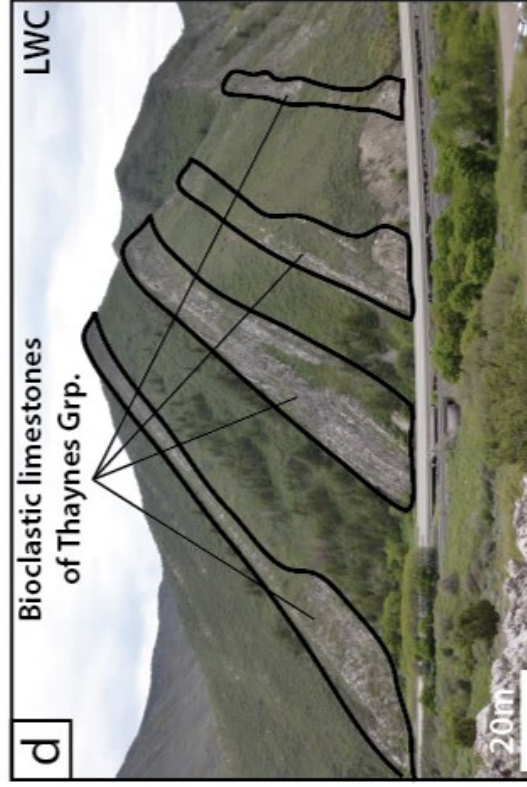
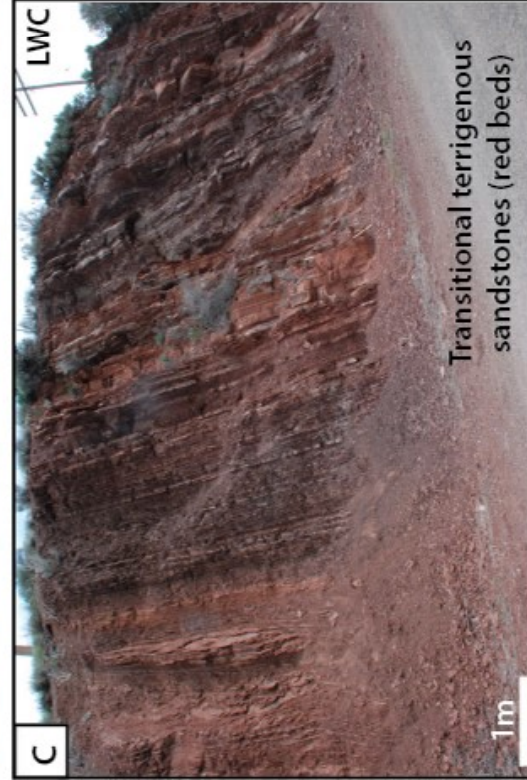
Main lithologies encountered in the basin:

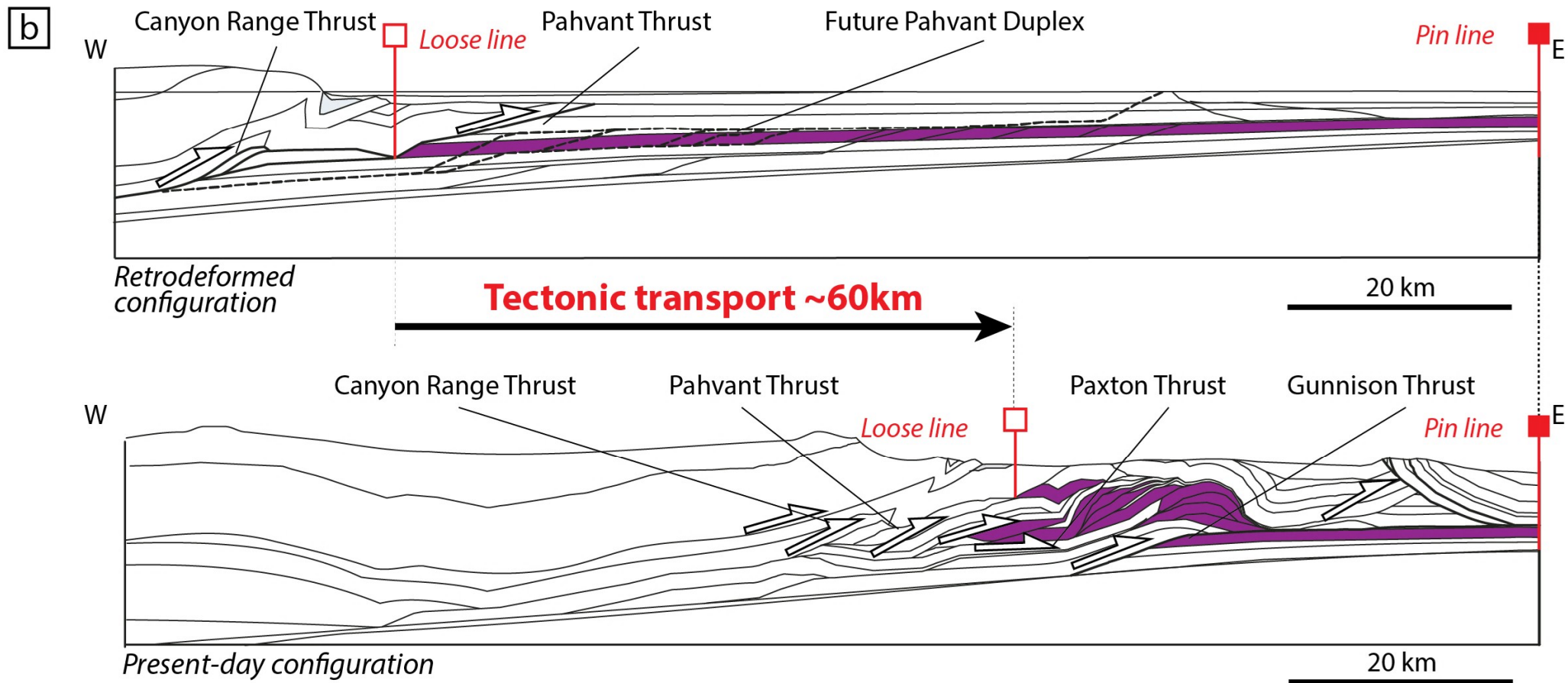
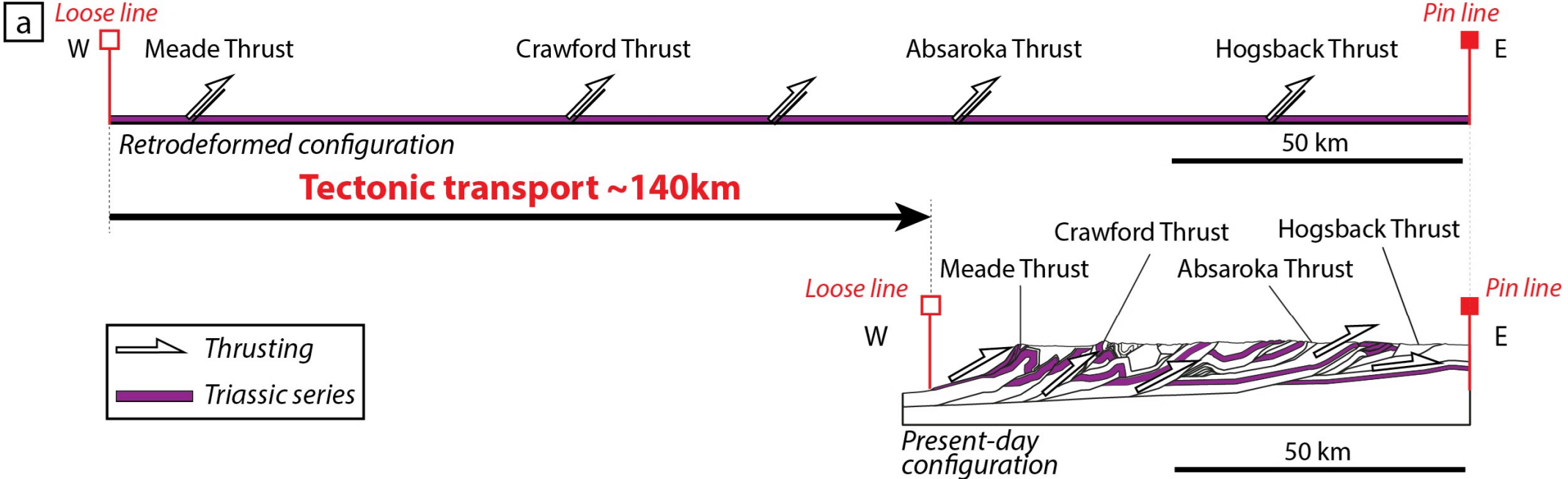
-  Marine siltstones
-  Marine shales
-  Marine bioclastic limestones
-  Intertidal microbial carbonates
-  Transitional terrigenous sandstones
-  Transitional terrigenous conglomerates

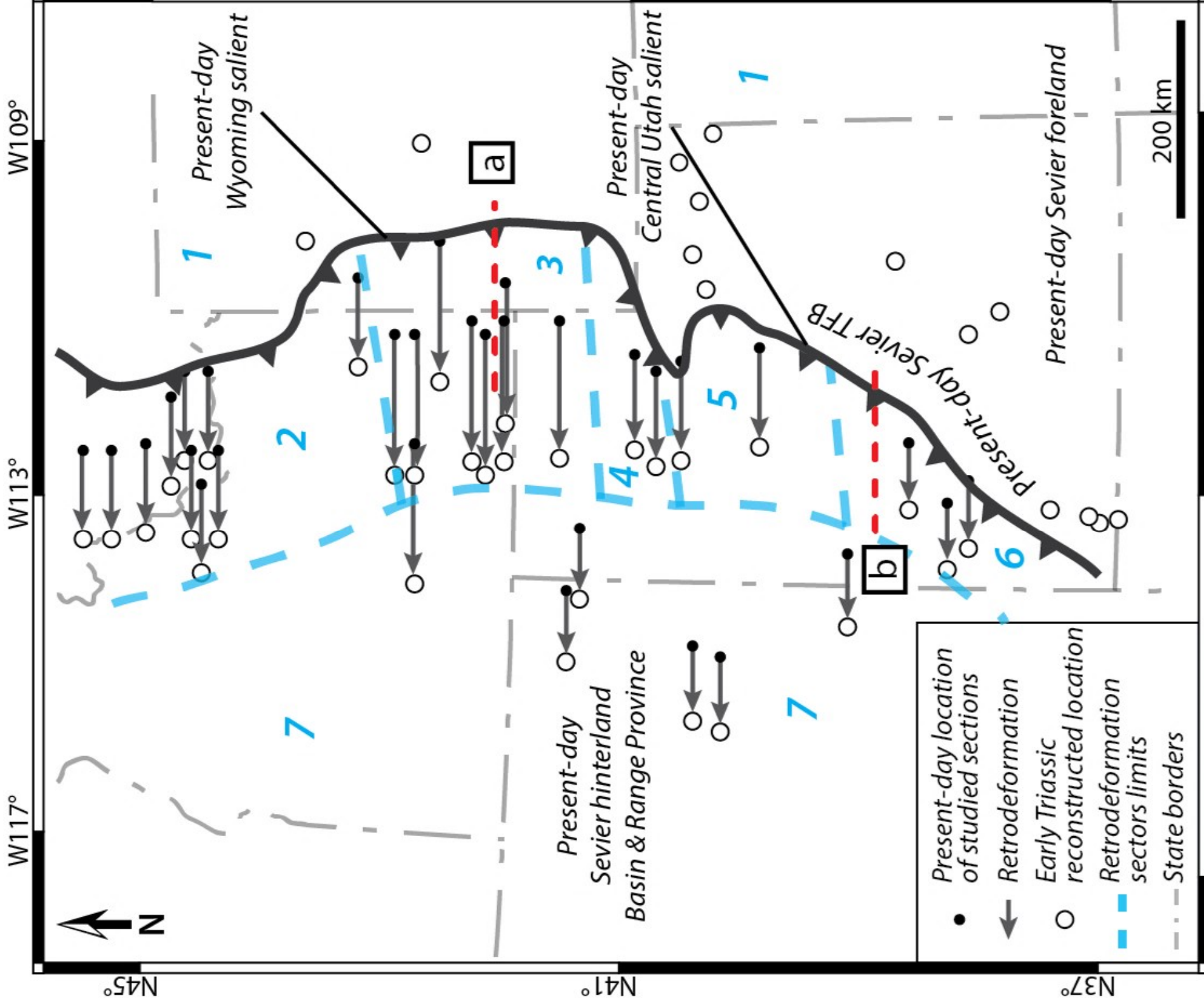
Main biostratigraphic markers:

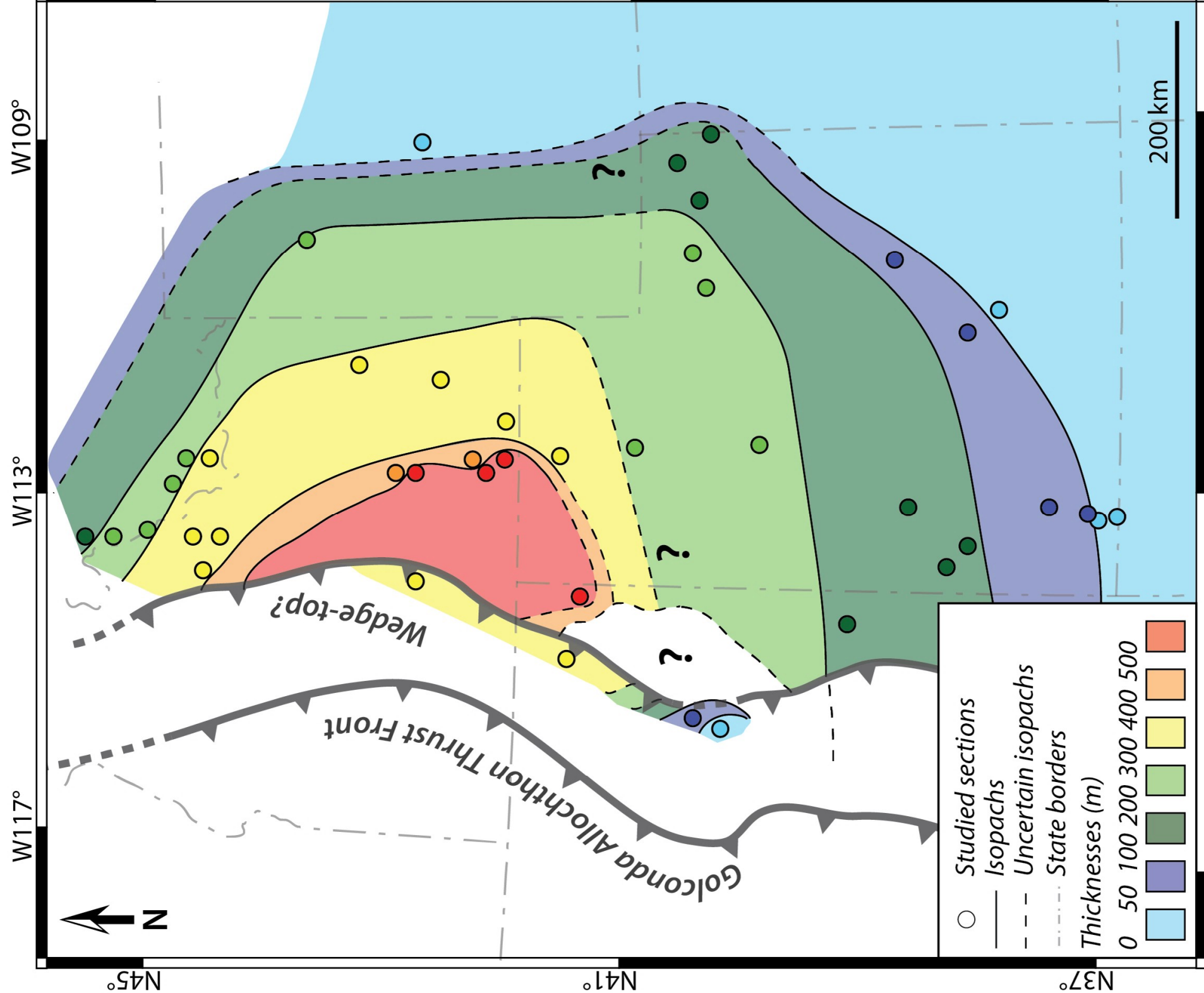
- Ana. Anasibirites beds
- Col. Columbites beds

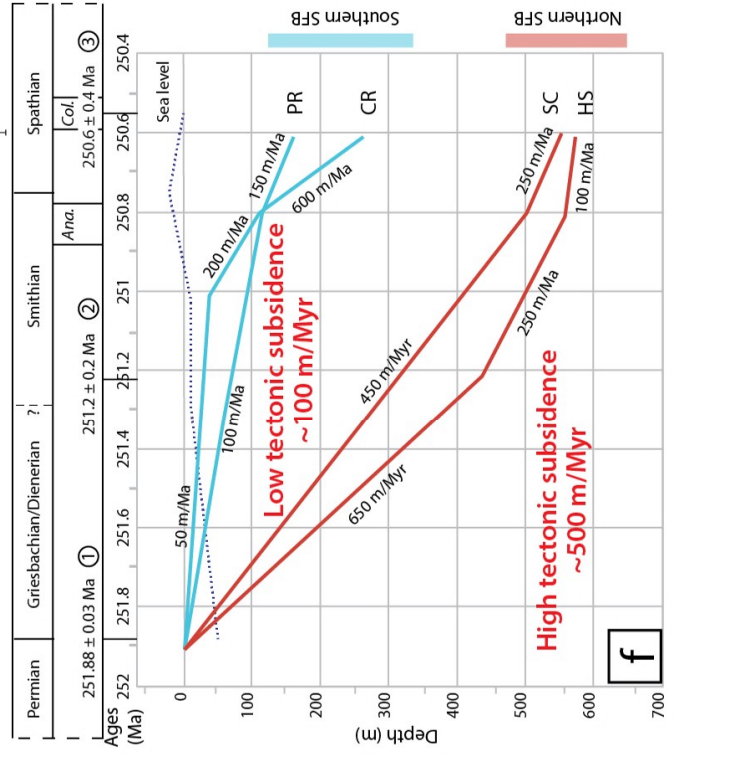
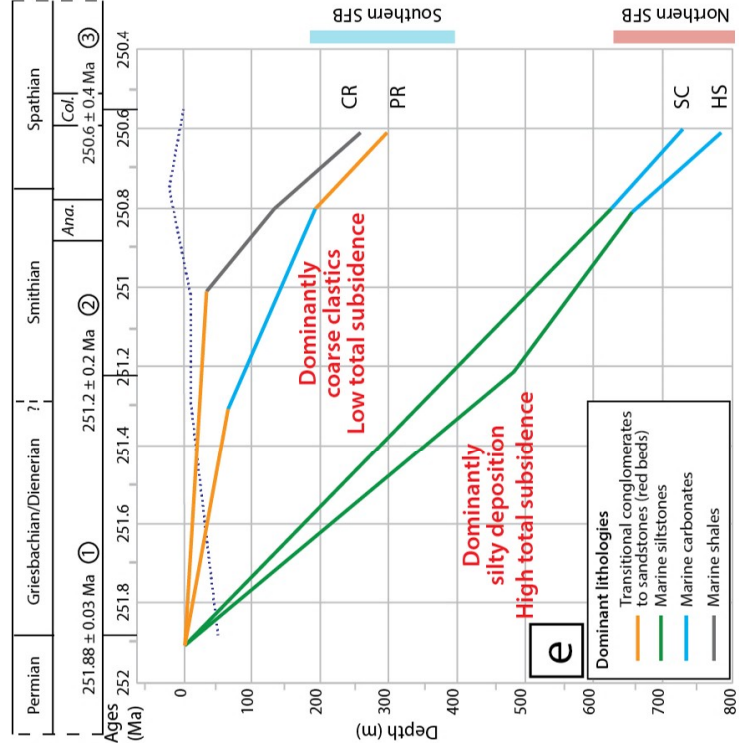
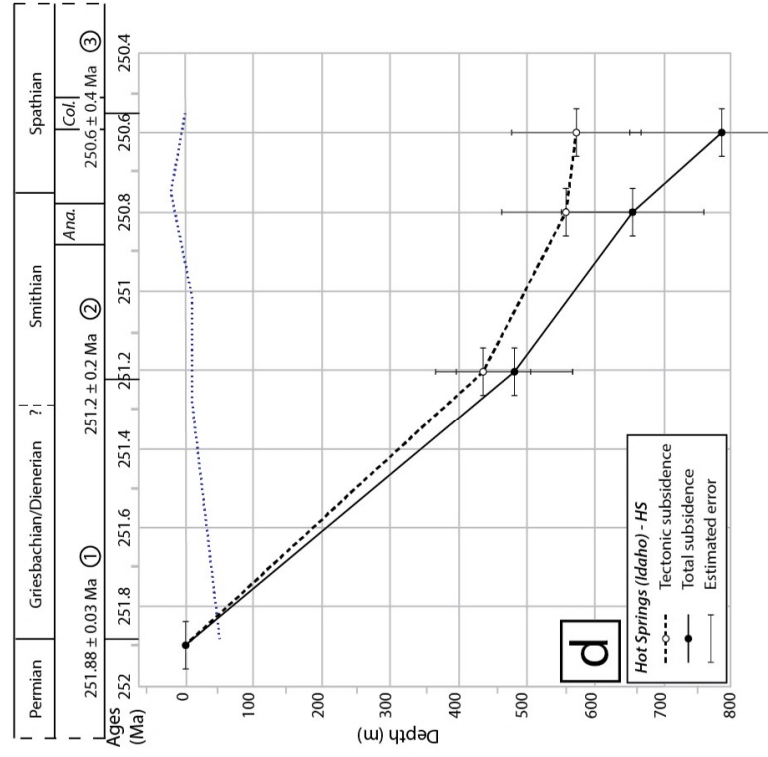
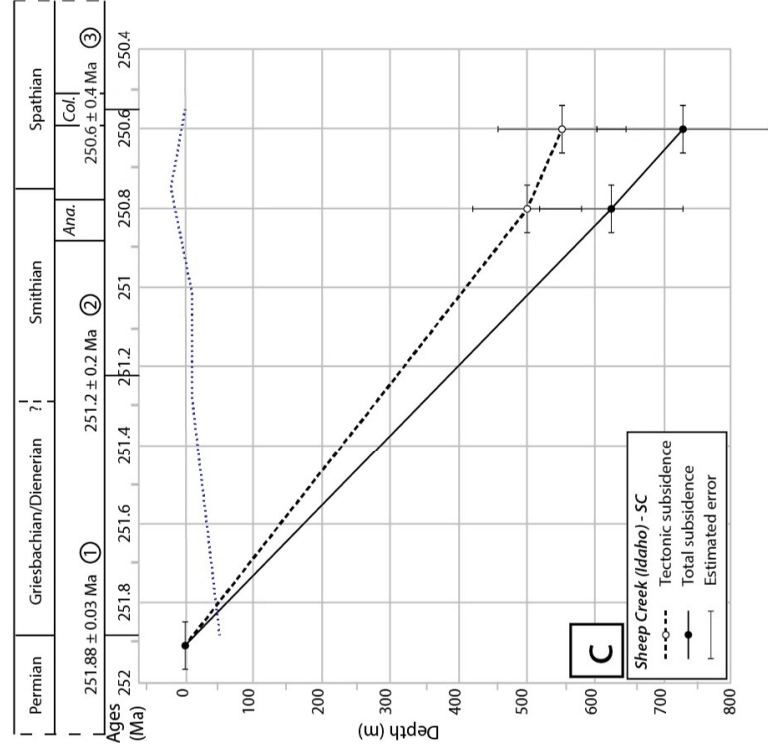
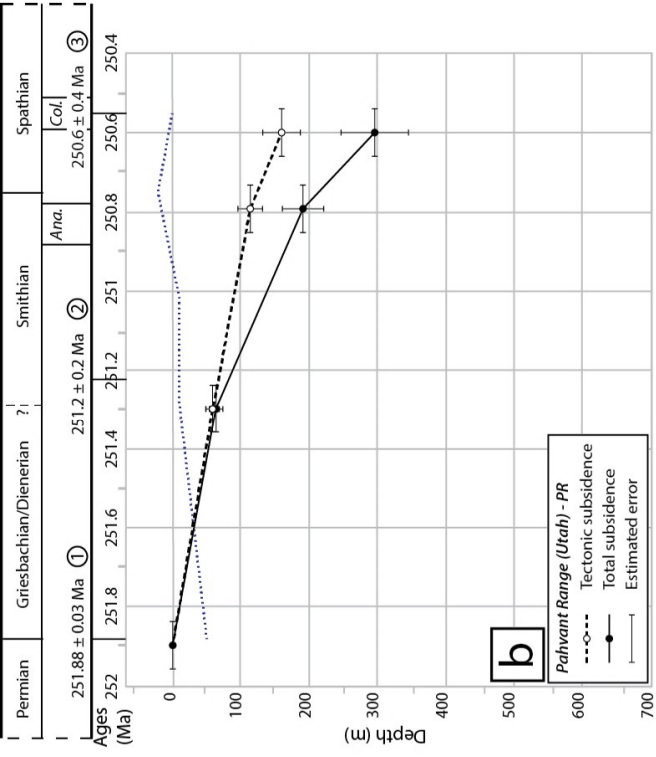
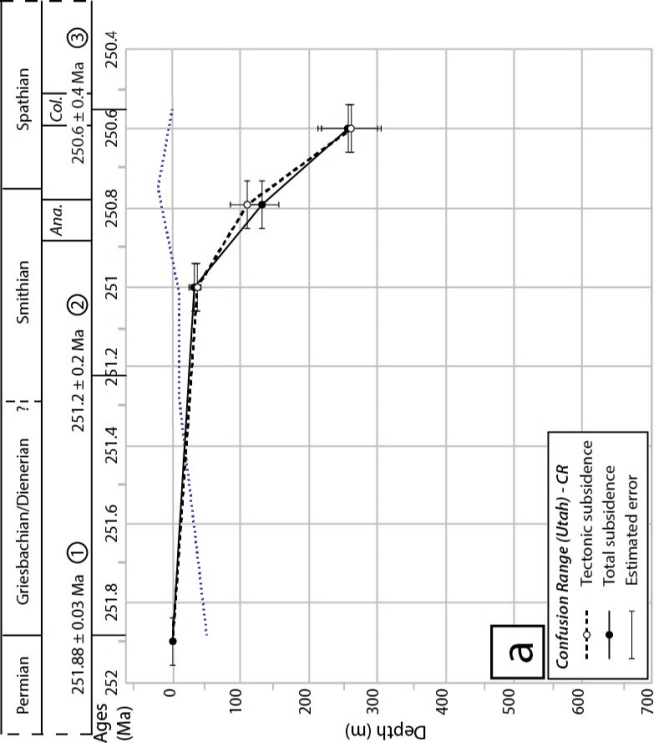


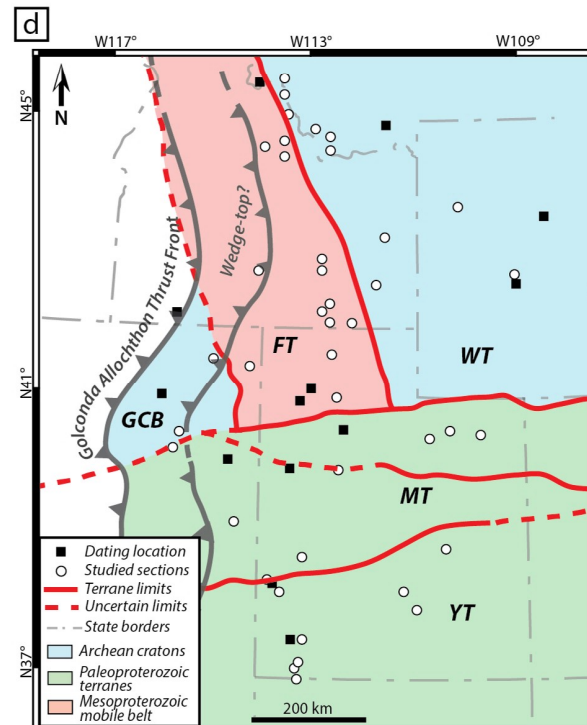
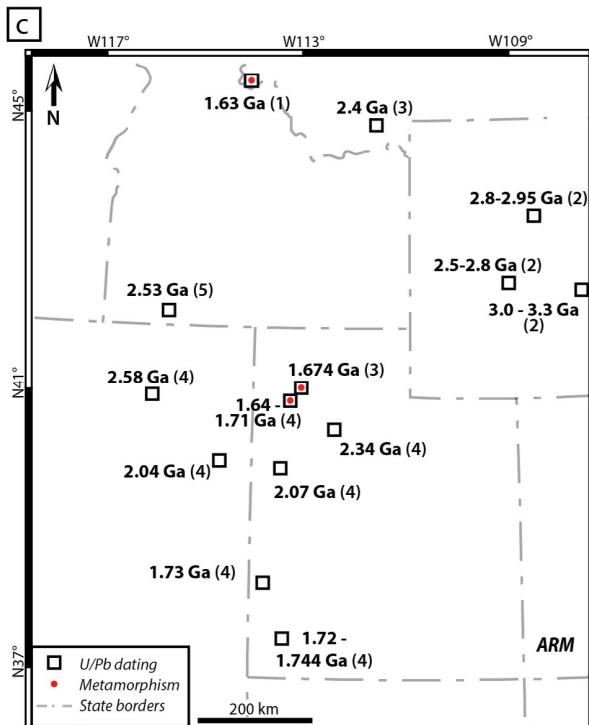
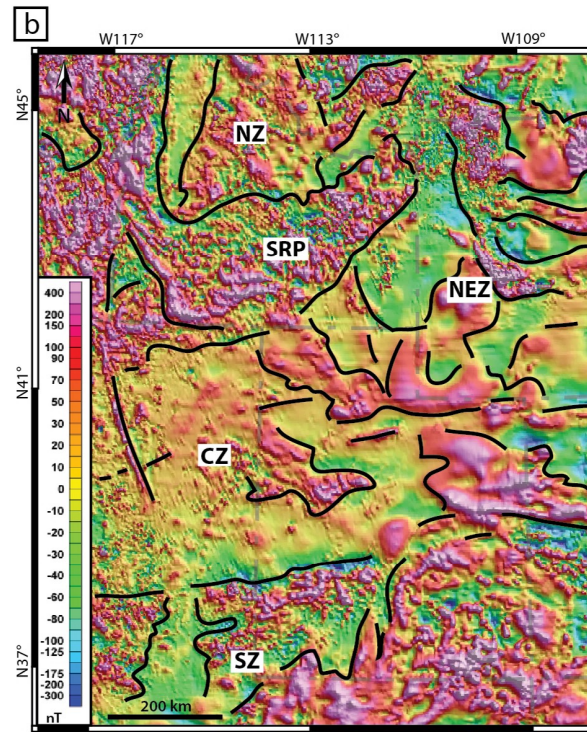
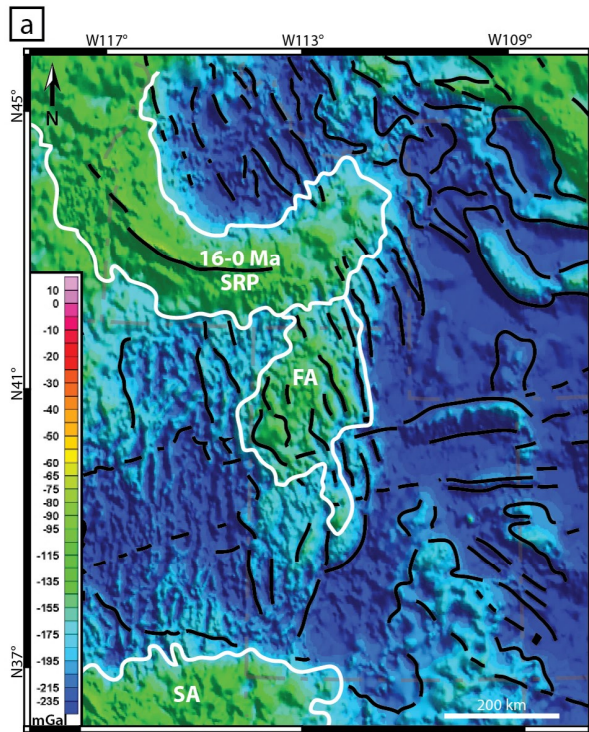


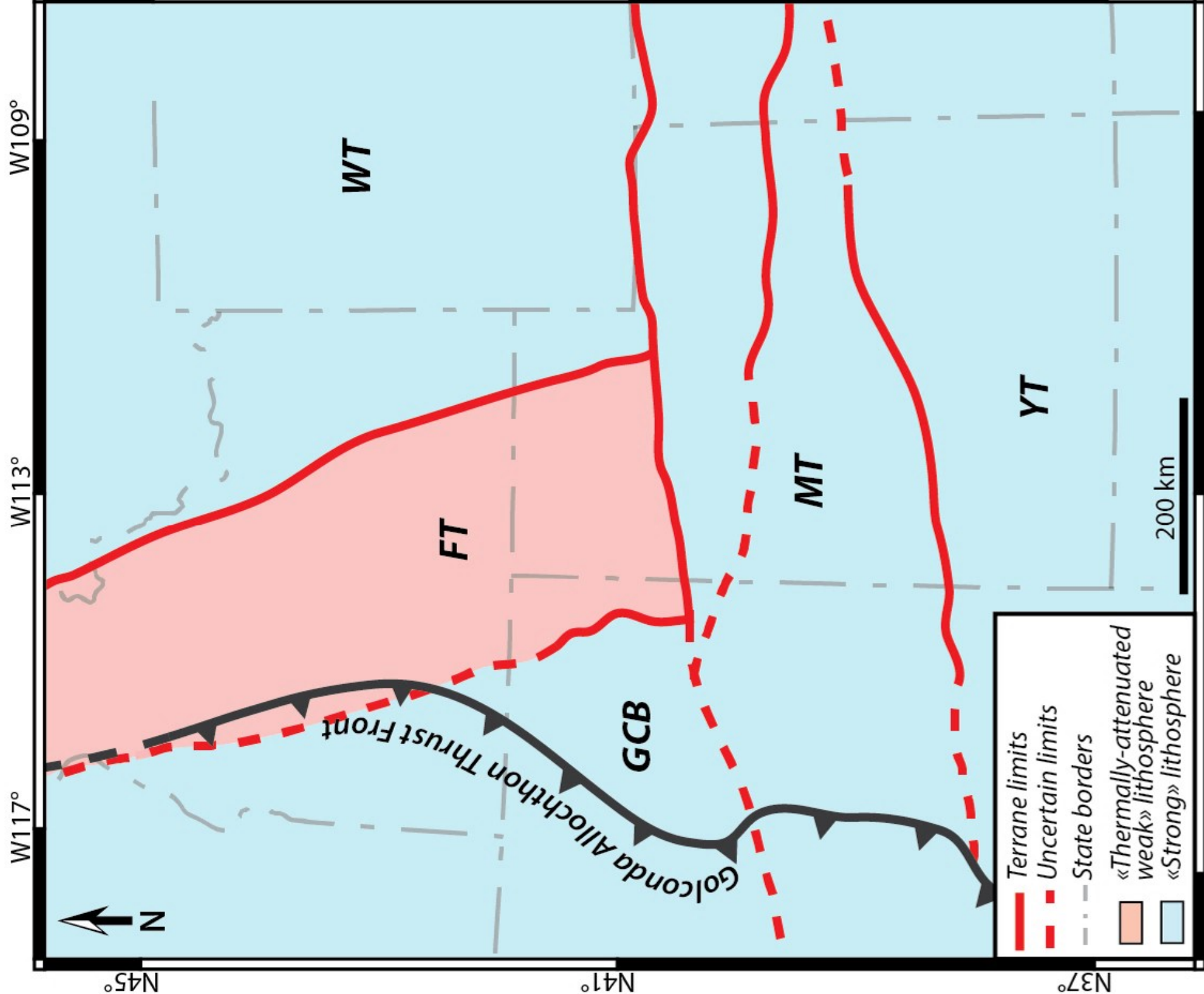


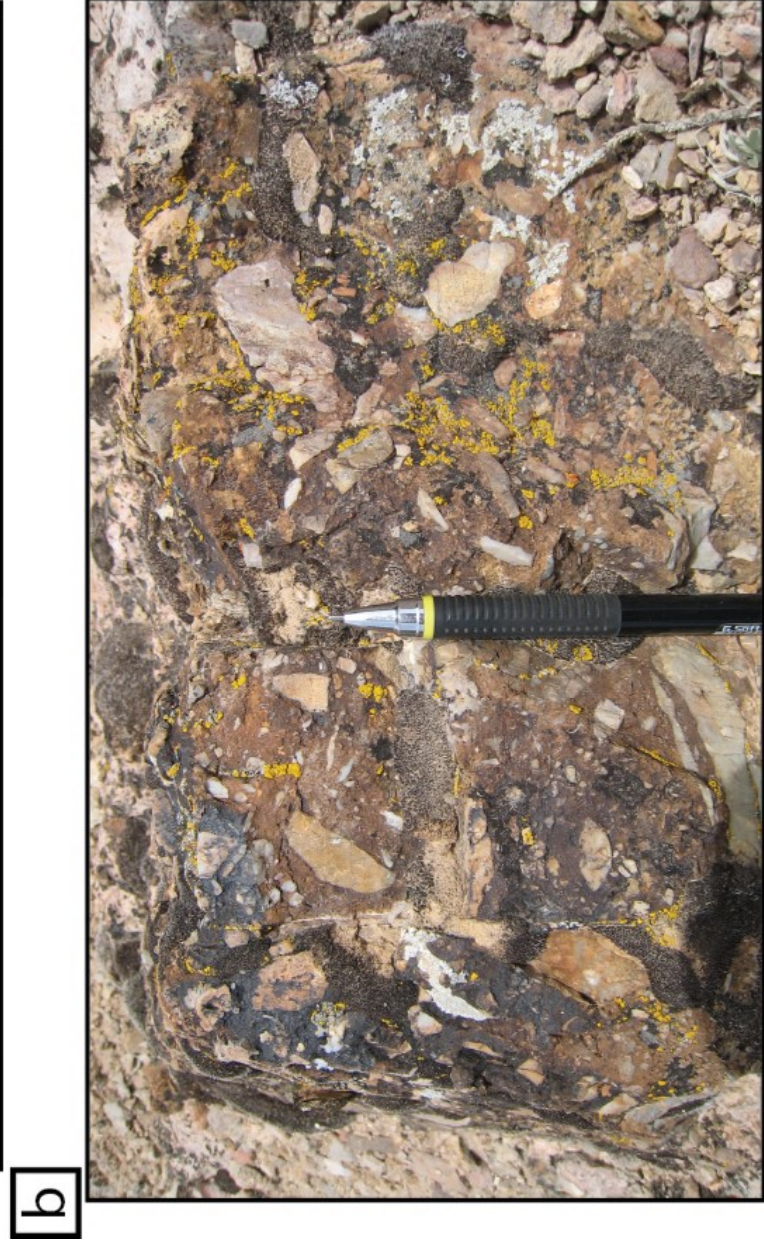
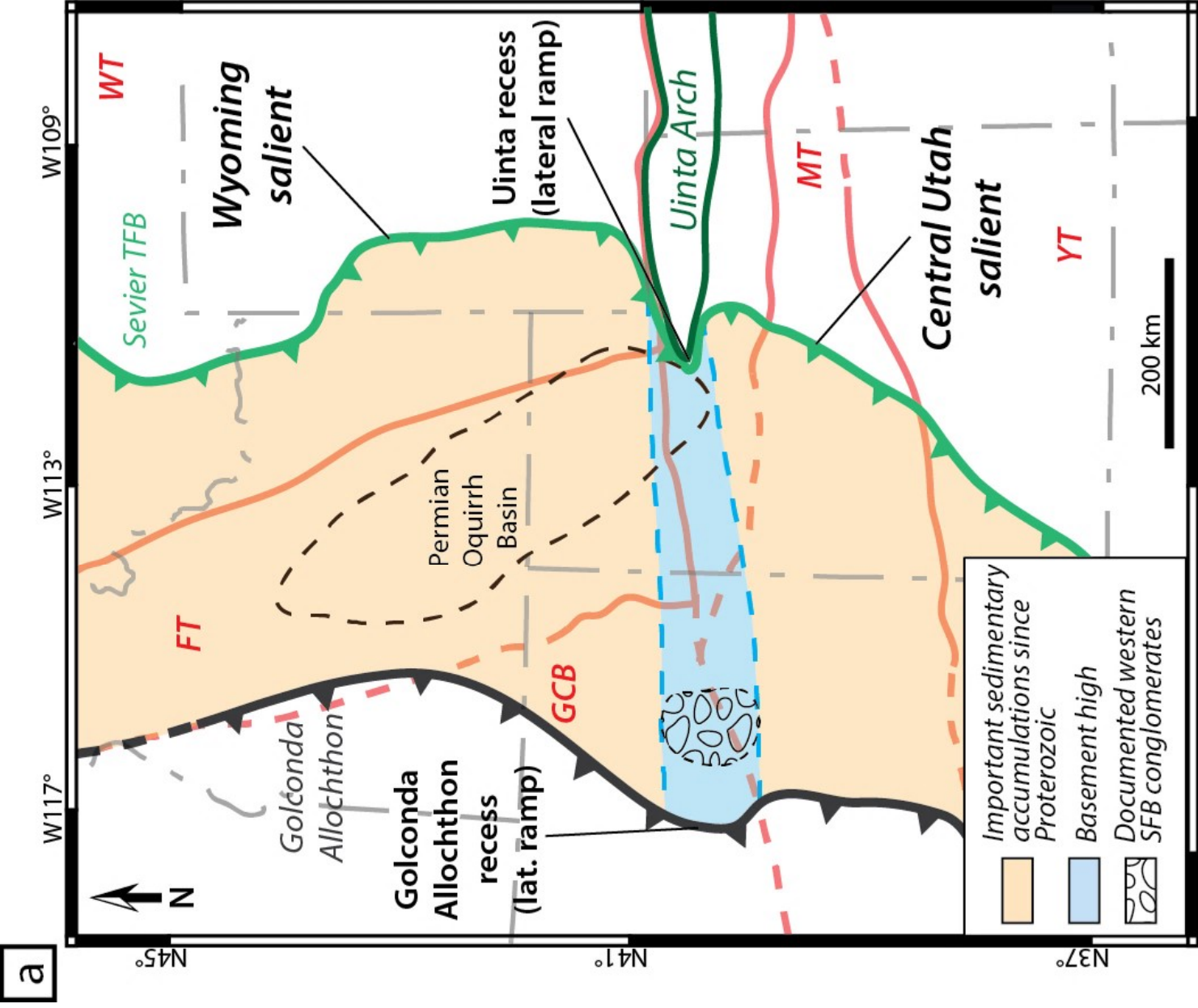


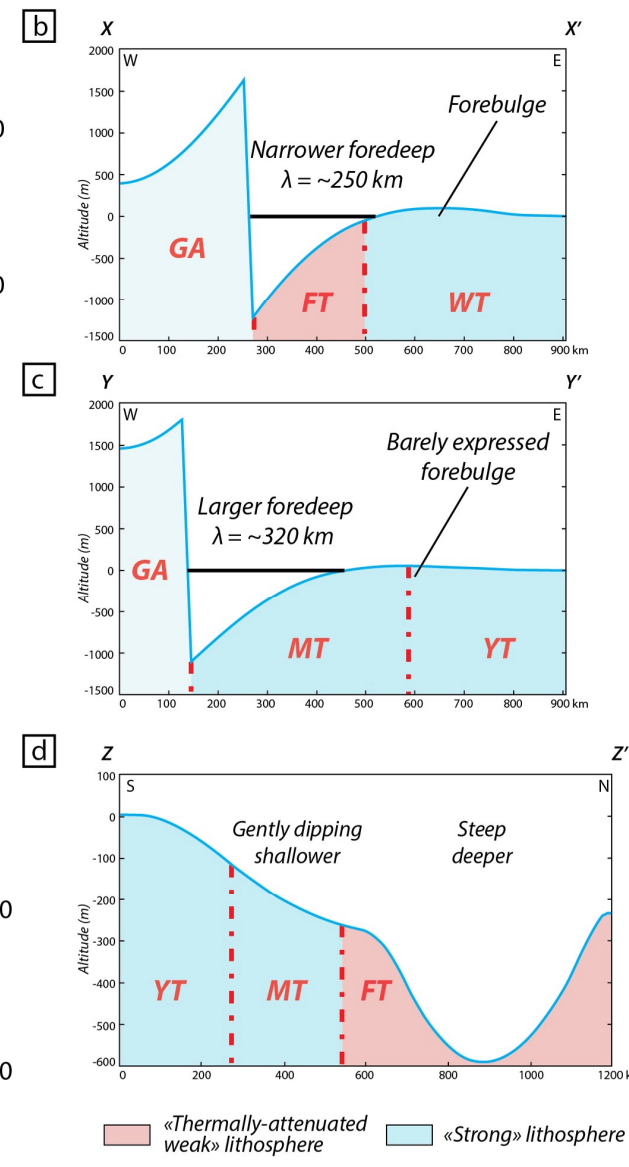
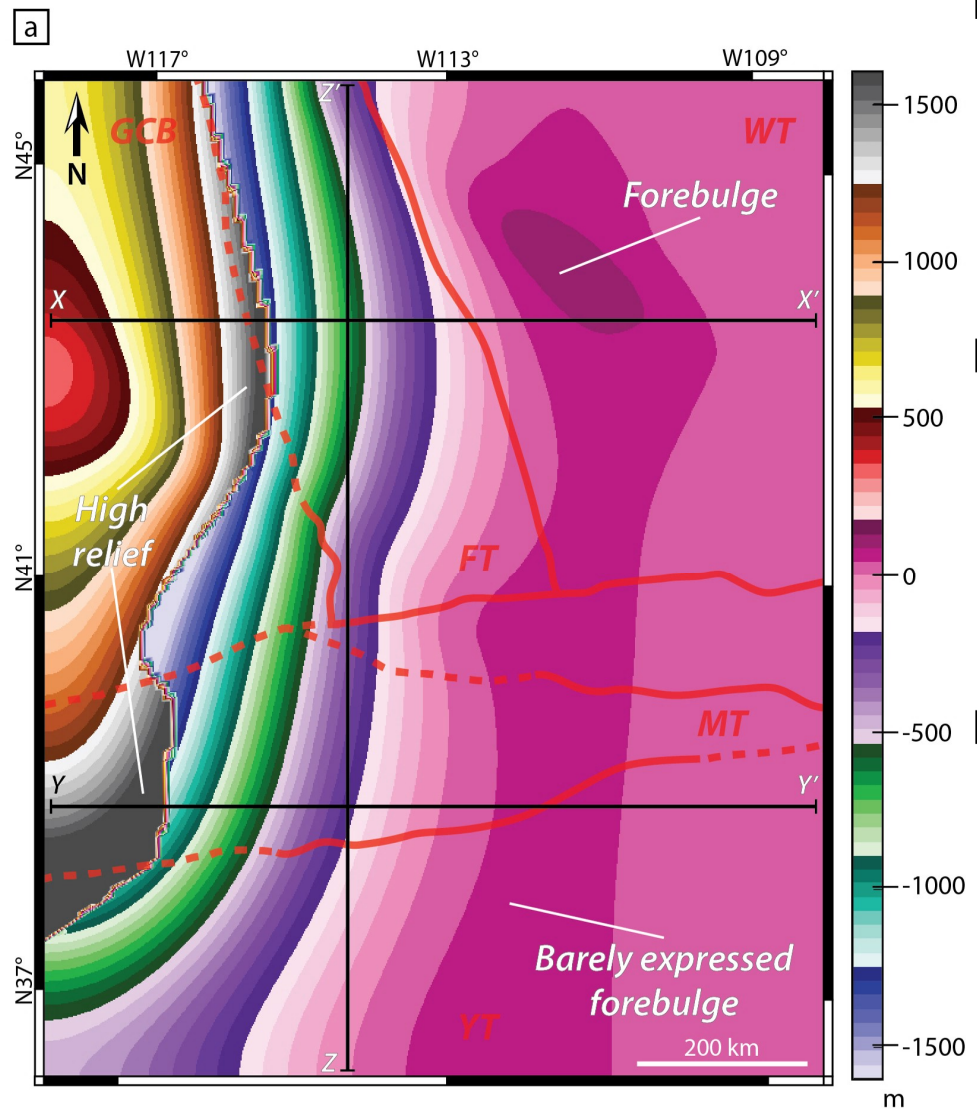






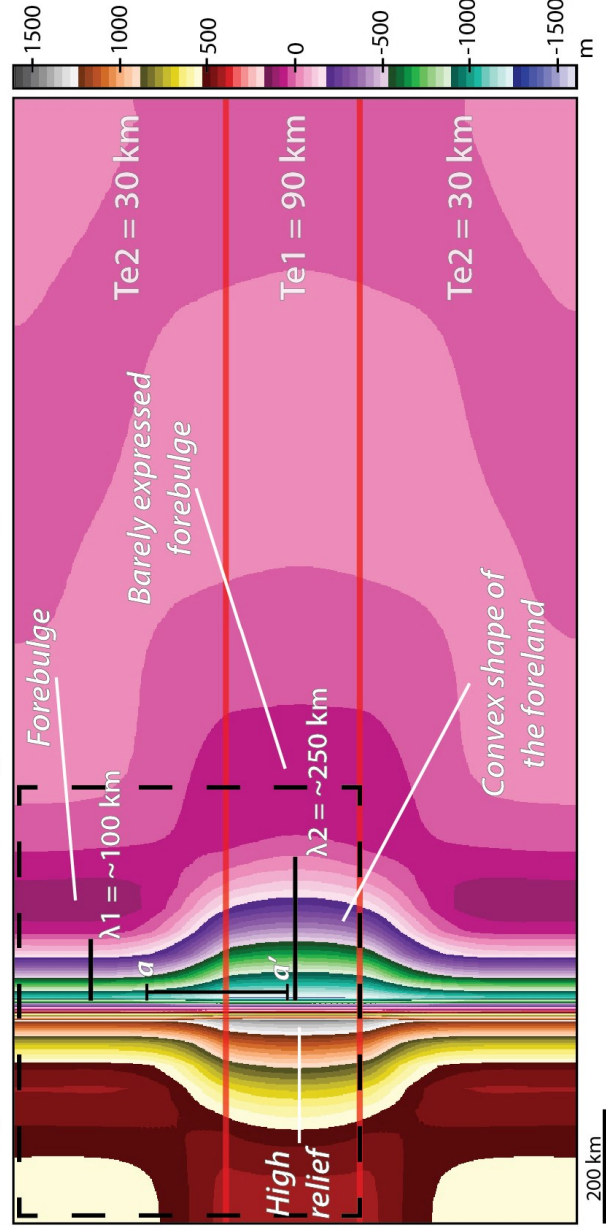






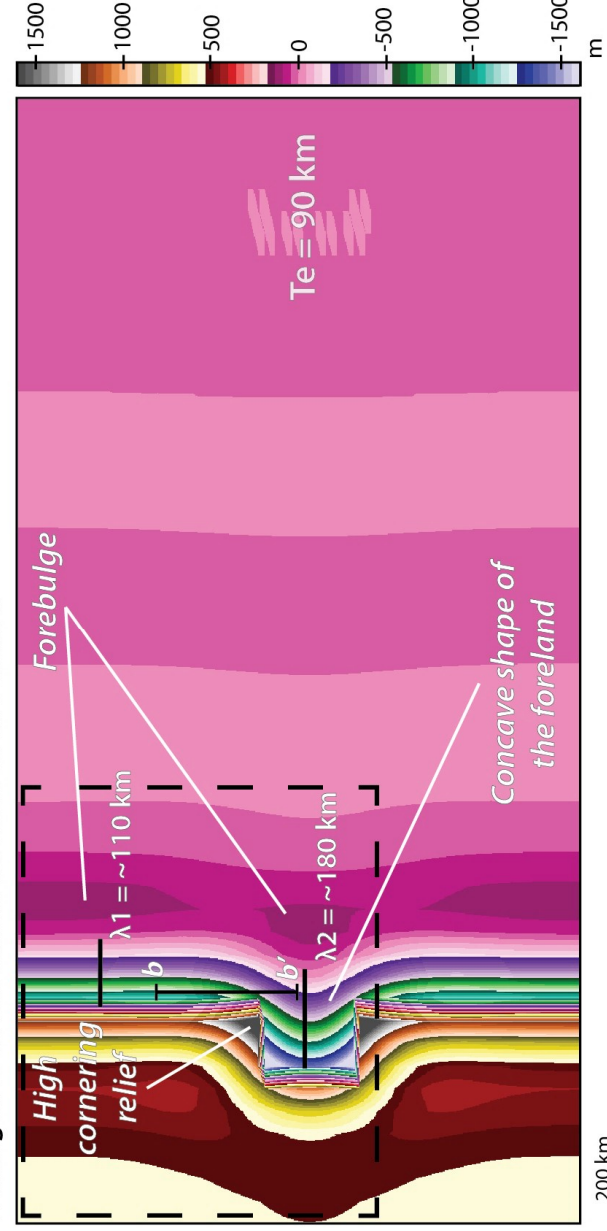
a

Heterogeneous basement & homogeneous allochthon



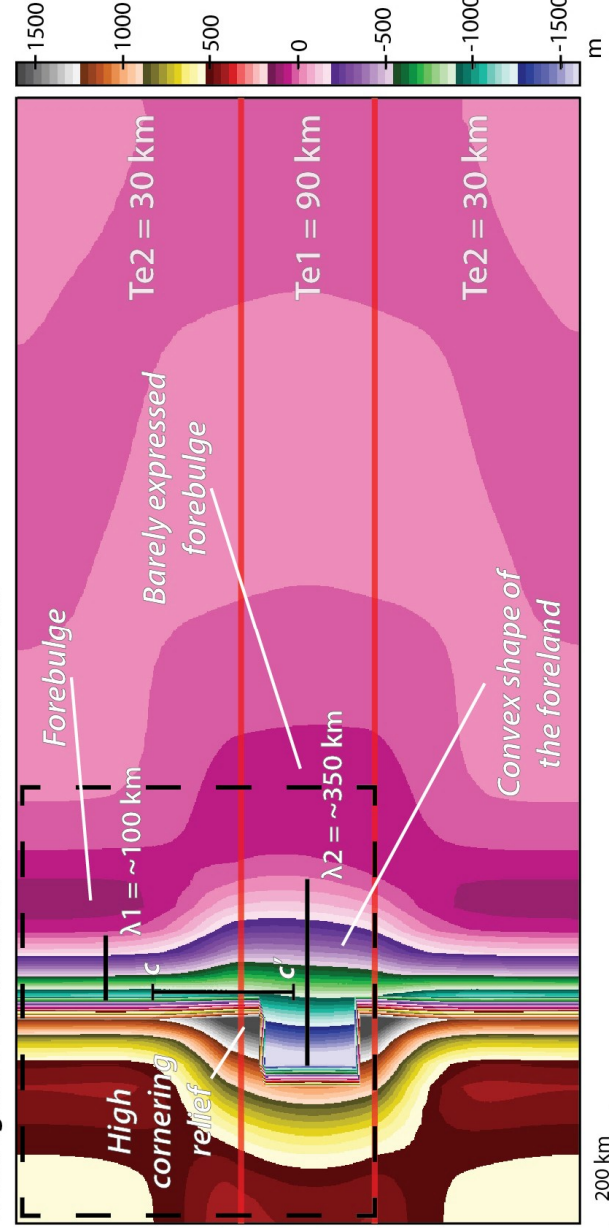
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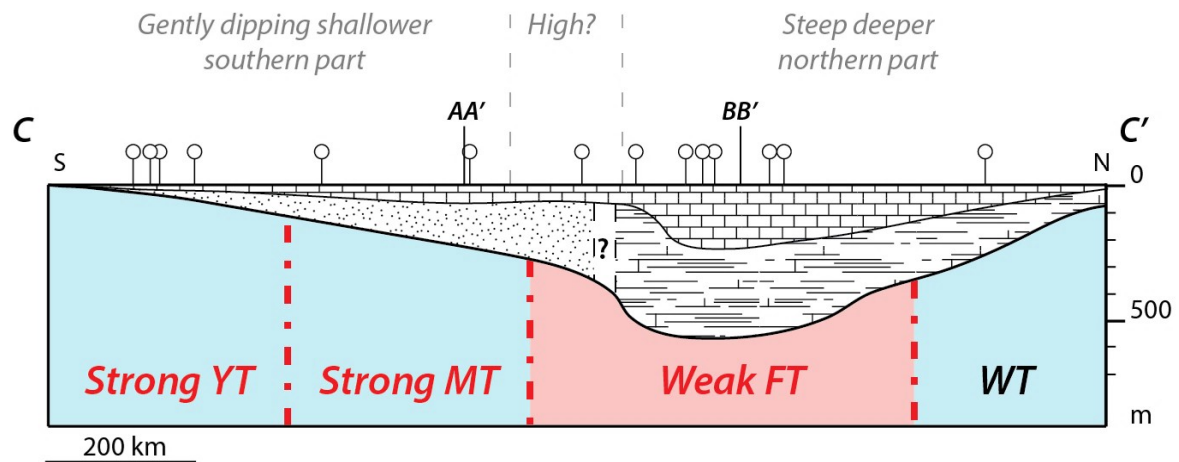
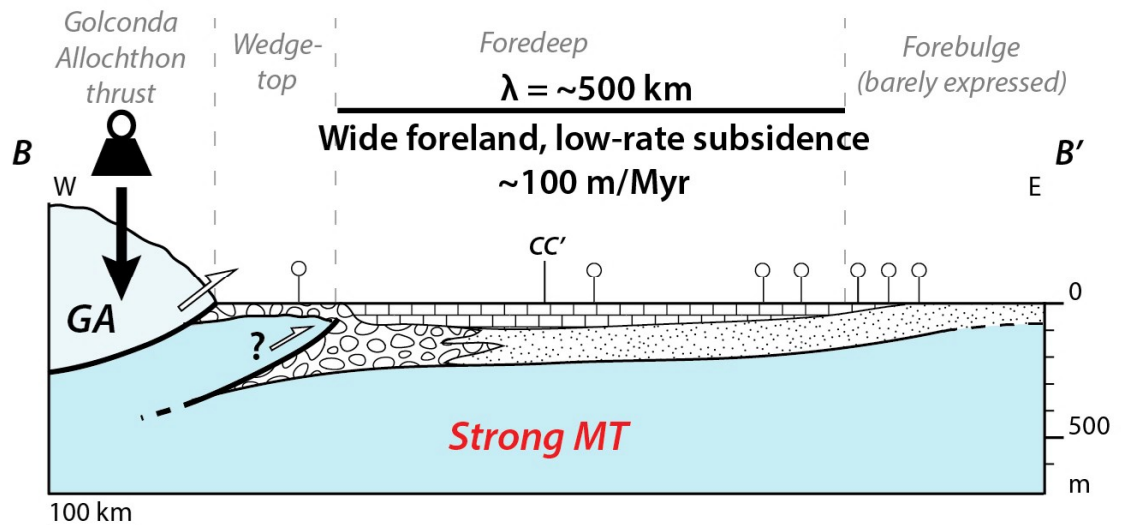
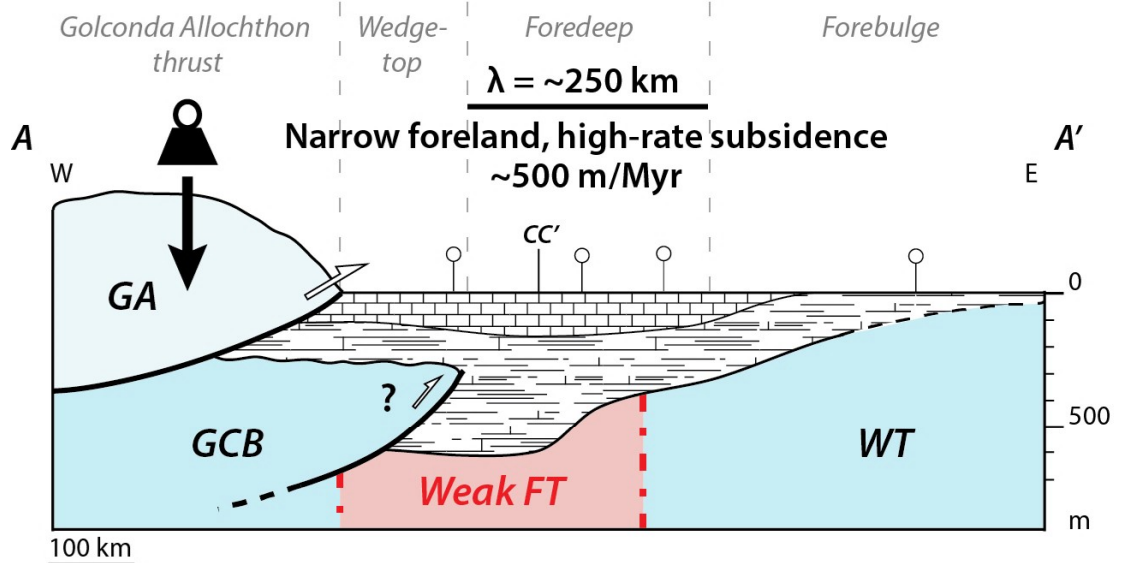
Homogeneous basement & recessed allochthon



c

Heterogeneous basement & recessed allochthon





SFB basement rheology

- «Strong» lithosphere
- «Thermally-attenuated weak» lithosphere

PTU-Smithian sediments dominant lithologies

- Marine siltstones
- Marine bioclastic limestones
- Transitional terrigenous sandstones
- Transitional terrigenous conglomerates

- Thrusting
- Golconda Allochthon topographic load
- Studied sections (see Figs. 3-8)
- Golconda Allochthon thrust front
- Postulated wedge-top

