

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

Gwénaël Caravaca, Arnaud Brayard, Emmanuelle Vennin, Michel Guiraud, Laetitia Le Pourhiet, Anne-Sabine Grosjean, Christophe Thomazo, Nicolas Olivier, Emmanuel Fara, Gilles Escarguel, et al.

▶ To cite this version:

Gwénaël Caravaca, Arnaud Brayard, Emmanuelle Vennin, Michel Guiraud, Laetitia Le Pourhiet, et al.. Controlling factors for differential subsidence in the Sonoma Foreland Basin (Early Triassic, western USA). Geological Magazine, 2018, 155 (6), pp.1305 - 1329. 10.1017/S0016756817000164 . hal-01848947

HAL Id: hal-01848947 https://hal.science/hal-01848947

Submitted on 28 Jan 2022

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers. L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

2	Foreland Basin (Early Triassic, western USA)				
3					
4	Published in <i>Geological Magazine</i> 155, 1305 - 1329				
5					
6	Gwénaël Caravaca ¹ , Arnaud Brayard ¹ , Emmanuelle Vennin ¹ , Michel Guiraud ¹ , Laetitia Le Pourhiet ² ,				
7	Anne-Sabine Grosjean ¹ , Christophe Thomazo ¹ , Nicolas Olivier ³ , Emmanuel Fara ¹ , Gilles Escarguel ⁴				
8	Kevin G. Bylund ⁵ , James F. Jenks ⁶ , Daniel A. Stephen ⁷ .				
9	¹ Biogéosciences UMR6282, CNRS, Univ. Bourgogne Franche-Comté, 21000 Dijon, France				
10	² Sorbonne Universités, UPMC Univ Paris 06, CNRS, Institut des Sciences de la Terre de Paris (iSTeP), 4				
11	place Jussieu 75005 Paris, France				
12	³ Laboratoire Magmas et Volcans, CNRS, IRD, OPGC, Université Blaise Pascal, 63038 Clermont				
13	Ferrand, France				
14	⁴ UMR 5023 LEHNA, Université Lyon 1, 69622 Villeurbanne Cedex, France				
15	⁵ 140 South 700 East, Spanish Fork, Utah 84660, USA				
16	⁶ 1134 Johnson Ridge Lane, West Jordan, Utah 84084, USA				
17	⁷ Department of Earth Science, Utah Valley University, Orem, Utah 84058, USA				
18	Contact details for corresponding author: Phone: +33.3.80.39.62.54. E-mail: gwenael.caravaca@u-				
19	bourgogne.fr				

Controlling factors for differential subsidence in the Sonoma

20 Abstract –

21 Sediments deposited from the Permian–Triassic boundary (~252<5>Ma) until the end-Smithian (Early Triassic; c. 250.7<5>Ma) in the Sonoma Foreland Basin show marked thickness variations between its 22 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

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

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

53Allochthon 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

The Sonoma Foreland Basin lies within a region of the North American continent 60 showing a very long and complex tectonosedimentary history starting during Proterozoic 61 time and still active today (e.g. Dickinson, 2013). The first documented structuring of the 62 region dates back to the Palaeoproterozoic period when Mojave and Yavapai terranes were 63 emplaced against the Archean Wyoming craton (Fig. 1b; Whitmeyer & Karlstrom, 2007, fig. 64 65 9; Lund et al. 2015). This event generated multiple crustal fault zones along which later reactivations were possible with deformational episodes (Oldow et al. 1989; Dickerson, 66 67 2003). At least two rifting events took place in this region during subsequent Proterozoic times (Burchfiel & Davis, 1975; Oldow et al. 1989), the most recent being during 68 Neoproterozoic time (c. 770<5>Ma) and linked to the fragmentation of the supercontinent 69 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<5>Ma) and corresponds to the deposition of a thick sedimentary prism formerly known as the 72 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 characterized by the emplacement of a large obducted accretionary prism located in Central 78 79 Nevada today (i.e. Roberts Mountains Thrust, Fig. 1c; Burchfiel & Davis, 1975; Speed & Sleep, 1982; Speed & Silberling, 1989; Burchfiel & Royden, 1991). The Roberts Mountains 80 81 Allochthon led to the formation of the N-S-trending westwards-dipping Antler Foreland

Basin (Speed & Sleep, 1982; Burchfiel & Royden, 1991; Blakey, 2008; Ingersoll, 2008;
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 early-middle Permian time (c. 350-270<5>Ma; Fig. 1b). This mountain-building event 86 resulted from a succession of crustal uplifts because of important long-range intracratonic 87 deformations. There, transtensional and transpressional constraints occurred along with 88 lithospheric buckling as a response to the Laurentia-Gondwana continental collision (Kluth 89 & Coney, 1981; Ye et al. 1996; Geslin, 1998; Dickerson, 2003; Dickinson, 2006, 2013; 90 Blakey, 2008). The resulting chain probably showed a marked topographic relief, some of 91 which could have persisted until Early Triassic time (Kluth & Coney, 1981; Blakey, 2008). 92 93 Most of these crustal uplifts were emplaced according to lithospheric weaknesses inherited from the Proterozoic structural events (Kluth & Coney, 1981; Dickerson, 2003). 94

Many sedimentary basins formed during the Carboniferous–Permian interval
(Dickerson, 2003). For instance, the Permian Oquirrh Basin (Fig. 1c) probably resulted from
the complex interplay between intracratonic deformations to the east and the reactivation of
Antler faults to the west (Geslin 1998: fig. 12; Trexler & Nitchman, 1990; Dickerson, 2003;
Blakey, 2008). This highly subsiding basin recorded up to 6<5>km of marine strata (Walker,
1985; Yonkee & Weil, 2015).

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

part of the basin, which are presently located in Central Nevada (Fig. 1c; Snyder & 112 Brueckner, 1983; Walker, 1985; Schweickert & Lahren, 1987; Oldow et al. 1989; Dickinson, 113 2006, 2013; Blakey, 2008; Ingersoll, 2008). Remnants of the Golconda Allochthon, if any, 114 are yet to be found in the northern part of the basin, especially in Idaho (Schweickert & 115 116 Lahren, 1987; Oldow et al. 1989). This allochthon is sealed in present-day Nevada by the rhyolitic Koipato Formation volcanism, presumably emplaced by the end of the Sonoma 117 orogeny (Vetz, 2011). A minimum age of Anisian (Middle Triassic) can be given to this 118 volcanic formation using geochronology (Vetz, 2011) and due to the occurrence of Anisian 119 120 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 discarded. 122

The following Sevier orogeny is of Early Cretaceous – Eocene age (c. 140–50<5>Ma; 123 Fig. 1b) and it originated from the subduction of the Farallon Plate under the North American 124 125 continental plate (Burchfiel & Davis, 1975; Dickinson, 2006, 2013). E-W-directed compressive constraints resulted in the formation of a large Sevier thrust-and-fold belt which 126 is still present today and constitutes the eastern border of the Great Basin (Fig. 1c; Dickinson, 127 2006, 2013; Yonkee & Weil, 2010; Yonkee et al. 2014). This thrust-and-fold belt is however 128 not homogeneous along its N-S-trending front, and displays two convex-to-the-foreland 129 'salients' (Fig. 2) with varying estimated tectonic shortening and eastwards displacement of 130 terrains reaching up to 140<5>km (DeCelles & Coogan, 2006; Schelling et al. 2007; 131 Dickinson, 2006, 2013; Yonkee & Weil, 2010, 2015; Yonkee et al. 2014). These Wyoming 132 and Central Utah salients are separated by a conspicuous recess formed by a lateral ramp and 133 134 located west of the Uinta Mountains (Figs 1c, 2). Its formation results from inherited features of the basement (see Section 4.c; e.g. Lawton, Boyer & Schmitt, 1994; Mukul & Mitra, 1998; 135 Paulsen & Marshak, 1999; Wilkerson, Apotria & Farid, 2002). 136

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

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

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

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

203 3.b. Methods

3.b.1. Palinspastic reconstructions using retrodeformations

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

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

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

retrodeformation of regional cross-sections taken from geological maps (cross-sections inFig. 6).

3.b.2. Subsidence analysis and backstripping

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

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

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

- radiochronological data, leading to a detailed timeframe with a maximum error of around
 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
- were integrated in Global Mapper v.16.2.3 GIS software (available at
- http://www.bluemarblegeo.com/products/global-mapper.php) to generate an isopach map by
- creating a 3D triangulated grid projection of thicknesses (online Supplementary Figure S1).
- 298 3.b.4. Lithospheric heterogeneity of the basement

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

310 3.b.5. Numerical model

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

319 4. Results

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

4.a. Lithological and stratigraphical analyses

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

Based on the palinspastic map, we constructed a palaeogeographic isopach map of the 335 SFB (Fig. 8). The isopach map shows that the distribution of the sedimentary thickness for 336 the PTU-Smithian interval is heterogeneous within the basin, showing a thicker succession in 337 338 the northern than in the southern part. In the southern part, the thickness gradually varies along a roughly NW-SE-aligned transect, showing low thicknesses over a large surface 339 (c. 500<5>km from east to west). The thickness ranges from a few tenths of metres in south 340 and SE Utah, up to 250<5>m around Salt Lake City. The westernmost area (NE Nevada) is 341 also characterized by low thicknesses (<100<5>m thick). Conversely, the northern part of the 342 basin exhibits a marked transition with thickness values broadly increasing from east to west. 343 The easternmost area of the northern part (west Wyoming) shows sedimentary thicknesses 344 similar to that of the southern part (<300<5>m thick; Fig. 8). The west-central area records 345 the thickest succession of the SFB (up to c. 550 < 5 > m thick), and is centred on present-day 346 south-central Idaho. The westernmost area (west-central Idaho) shows similar thicknesses (up 347 to c. 300<5>m thick; Fig. 8). 348

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

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

362 Finally, the tectonic subsidence appears as a critical diagnostic feature for the basin (Fig. 9f). A marked difference exists between mean tectonic subsidence rates in the southern 363 and northern parts of the basin (c. 100<5>m Ma⁻¹ v. c. 500<5>m Ma⁻¹, respectively). The 364 southern sections show a low-rate tectonic subsidence ($50-200 < 5 > m Ma^{-1}$; Fig. 9e). 365 Nevertheless, a marked increase in subsidence rate is recorded during early Spathian time for 366 these sections (150–600<5>m Ma⁻¹; Fig. 9e). Conversely, the northern sections show a 367 higher rate of tectonic subsidence during the PTU-Smithian interval (450–650<5>m Ma⁻¹; 368 Fig. 9e), whereas early Spathian time is characterized by a decrease in subsidence rate (100– 369 370 250<5>m Ma⁻¹; Fig. 9e).

4.b. Basement characterization

On the gravimetric anomaly map shown on Figure 10a, black lines outline the geophysical features that may represent traces of crustal/lithospheric faults or heterogeneities in the basement (Lowrie, 2007). The lowest Bouguer anomaly values (<150<5>mGal, Fig. 10a) suggest the presence of a thick crust, whereas moderate negative anomalies (between -65 and -135<5>mGal; white outlines) point towards a thinner crust and/or the presence of lower-crustal high-density bodies (e.g. Gilbert, Velasco & Zandt, 2007; Lowrie,

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 the properties of the basement (Dickinson, 2013). The Farmington Anomaly (FA on 380 Fig. 10a), located in the centre of the study area, may result from the presence of lower-381 crustal high-density mafic and/or ultramafic material emplaced during a thermal event dated 382 383 at c. 1.64<5>Ga (Mueller et al. 2011). Alternatively, it can have originated from a more recent thermal event and/or the presence of a thin lithospheric crust (e.g. Gilbert, Velasco & 384 385 Zandt, 2007; Lowrie, 2007). Remnants of an important thermal metamorphism including partial melting (c. 1.67<5>Ga) can also be observed in this area (red dots in Fig. 10c; Mueller 386 387 et al. 2011). The Southern Anomaly (SA on Fig. 10a) is poorly documented and may result from variations in the crustal thickness of the terrane (e.g. Gilbert, Velasco & Zandt, 2007; 388 389 Lowrie, 2007), possibly linked to the Ancestral Rocky Mountains orogeny or to the more recent Laramide orogeny and the building of the Rocky Mountains (Ye et al. 1996; 390 391 Dickerson, 2003).

392 The aeromagnetic anomaly map presented in Figure 10b discriminates areas of contrasted magnetic signatures (separated by black lines on Fig. 10b). These disturbances in 393 magnetic field are attributed to differences in the nature of the rocks composing the basement 394 (Turner, Rasson & Reeves, 2007). We do not attempt to identify the exact nature of these 395 rocks here; rather, we use these contrasted anomalies to characterize differences of rock types 396 397 that compose the basement (Purucker & Whaler, 2007; Lund et al. 2015). As on the Bouguer gravity anomaly map, the presence of the Snake River Plane hotspot-track (SRP in Fig. 10a, 398 b) is obvious on the aeromagnetic anomaly map. It features a strong positive magnetic 399 anomaly signal (>150<5>nT, Fig. 10b). The Southern magnetic Zone (SZ on Fig. 10b) can be 400 401 distinguished on the southern part of the studied area by contrasted anomalies with a wide range of variations (from c. -200<5>nT up to c. 400<5>nT). The Central magnetic Zone (CZ 402 403 on Fig. 10b) occupies the central third of the map. It is characterized by generally neutral to (strongly) positive anomalies (from c. -10 < 5 > nT to c. 60 < 5 > nT, locally up to >150 < 5 > nT). 404 405 In the northeastern quarter of the studied area, the North-Eastern magnetic Zone (NEZ on Fig. 10b) is characterized by generally negative anomalies (between c. -80 < 5 > nT and c. -406 10<5>nT). Some areas with strong positive anomalies (>150<5>nT) are also observed, 407 whose shape and extension are very similar in the Bouguer gravity anomaly map (Fig. 10a). 408 409 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).

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

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 (MT); the Yavapai Terrane (YT); and the Farmington Terrane (FT; Fig. 10d). The GCB and 424 425 WT are Archean terranes with ages of c. 2.5<5>Ga (Nelson, Hart & Frost, 2011; Strickland, Miller & Wooden, 2011) and 2.4–3.3<5>Ga (Fan et al. 2011; Mueller et al. 2011), 426 427 respectively. The MT is a Palaeoproterozoic terrane of age 2.04–2.34<5>Ga, whereas the YT is a younger Palaeoproterozoic terrane of age 1.720-1.744<5>Ga (Nelson, Hart & Frost, 428 429 2011). The FT is a Mesoproterozoic intracratonic mobile belt (Lund et al. 2015) composed of reworked Archean crust (Whitmeyer & Karlstrom, 2007), with metamorphism ages between 430 431 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

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

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

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

Due to the lithospheric heterogeneity of the basement, the role of the boundary 454 lithospheric faults can be considered as essential. Neoarchean-Palaeoproterozoic terranes are 455 limited by mega-shear zones along with deep (nearly) vertical crustal and/or lithospheric 456 457 faults (Figs 10d, 11). Terranes in the SFB display some characteristics (e.g. dimension, geometry) that are similar to the terranes associated with the Neoarchean-Palaeoproterozoic 458 accretionary orogens (e.g. Chardon, Gapais & Cagnard, 2009, fig. 2; Cagnard, Barbey & 459 Gapais, 2011, fig. 9). These lithospheric and crustal accidents have therefore been reactivated 460 461 since their Precambrian onset (e.g. Bryant & Nichols, 1988; Paulsen & Marshak, 1999). Additionally, several authors (e.g. Eardley, 1939; Peterson, 1977) identified the presence of a 462 topographic basement highland (pale blue area in Fig. 12a, in colour online) near the junction 463 between the MT and the GCB/FT/WT during Palaeozoic time, separating the northern and 464 southern areas of marked sedimentary accumulation. Eardley (1939) first introduced this 465 feature as the 'Northern Utah Highland'. Peterson (1977) highlighted its presence on his 466 palinspastic maps for the Palaeozoic stratigraphic record. Finally, this sedimentary and 467 topographic pattern seems to have been the same in this basin since Proterozoic time (Paulsen 468 469 & 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 6<5>km of marine sediments accumulated in the Permian Oquirrh Basin in the northern part 472 of the SFB (Fig. 12a; Yonkee & Weil, 2015), whereas the southern part of the SFB saw the 473 474 deposition of only several hundred metres of marine and terrigenous sediments (e.g. c. 640<5>m in southwestern Utah; Rowley et al. 2005) during the same interval. The thick 475 476 Palaeozoic sedimentary series in northern and southern parts of the foreland (Peterson, 1977) would have allowed the thrust belt to propagate, while the presence of the topographic 477 478 basement highland characterized by a reduced sedimentary cover should have triggered the formation of a lateral ramp and a recess in the central part of the front (Fig. 12a). The 479 480 presence of the topographic high is attested by the occurrence of shallow conglomerates in the western part of the SFB within the PTU-Smithian interval (Fig. 12a, b; e.g. Gabrielse, 481 482 Snyder & Stewart, 1983; Lucas & Orchard, 2007; Jattiot et al. in press). Previous reconstruction of the GA thrust front also accounted for the presence of a recess in the central 483 part of the thrust front (e.g. Dickinson, 2006, 2013). Moreover, this mechanism underlying 484 the observed differential propagation has been proposed by Paulsen & Marshak (1999) for 485 the Sevier thrust-and-fold belt which shows the presence of a lateral ramp in its central part 486 (Fig. 2). This was explained by the pre-deformational sedimentary thicknesses pattern 487 showing thrusts propagating further when emplaced upon a thicker sedimentary cover 488 (Figs 2, 12a; Paulsen & Marshak, 1999, fig. 7). It is worth noting that both the lateral ramps 489 of the Sevier and Golconda thrust-and-fold belt are located close to and along the lithospheric 490 491 boundary between the MT and FT/WT (Figs 2, 12a).

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

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

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

499 4.d.1. Numerical approach and setup

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

502
$$\nabla^2 \left(D \nabla^2 \omega \right) = g \left(\rho_{\rm m} - \rho_{\rm i} \right) + q \tag{1}$$

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

506
$$D = \frac{ET_{\rm e}^{3}}{12(1-v^{2})},$$

depends solely on the effective elastic thickness T_e as the plate Young's modulus E and 507 Poisson's ratio v are fixed at 80<5>GPa and 0.25, respectively (Burov & Diament, 1995). 508 The topographic loads $q = \rho_t g h$ account for the thickening h resulting from the orogeny and 509 are computed using a density $\rho_{\rm f} = 2700 < 5 > \text{kg m}^{-3}$. The mantle restoring forces are computed 510 assuming a density $\rho_m = 3300 < 5 > \text{kg m}^{-3}$, while the infill is considered to be sediments of 511 density $\rho_1 = 1600 < 5 > \text{kg m}^{-3}$. We arbitrarily attributed a constant height h = 1500 < 5 > m to the 512 topographic load as we concentrate on the effect of heterogeneities of the allochthon 513 morphology and rheology of the basement only. These initial parameters are summarized in 514 Table 2. 515

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

In this model, we used $T_{e1} = 90 < 5 > \text{km}$ for the 'strong' GCB, WT, MT and YT lithospheres (Table 2), which is a good approximation for cratonic T_e (Watts, 1992). The 'weak-attenuated' FT is expected to show a contrasted lower T_e value due to its assumed rheological weaknesses. This value was set at $T_{e2} = 30 < 5 > \text{km}$ (Table 2; e.g. Leever *et al.* 2006).

528 4.d.2. Model results

529 Figure 13 shows that the southern part of the front is reconstructed as less propagated into the foreland than the northern part (Fig. 12a; see Dickinson, 2006, 2013). In this model, 530 the lateral ramp is spatially restricted along the limit between the FT/WT and MT (Fig. 13a). 531 The northern part, emplaced mainly above the 'weak' FT and in front of the largest part of 532 the GA, presents a narrower foredeep with $\lambda \approx 250 < 5 > \text{km}$ (Fig. 13a, b). The steep foredeep is 533 bordered by a well-expressed forebulge emplaced close to the FT/WT boundary (Fig. 13a; 534 XX' in Fig. 13b). The southern part of the foreland is set upon 'strong' lithospheres (MT and 535 YT) in front of the smallest and recessed parts of the GA (Fig. 13a, c). The foredeep in this 536 part of the model is larger, with $\lambda \approx 320 < 5 > \text{km}$, and its profile (YY' in Fig. 13c) also exhibits 537 a weaker topography than in the northern part. We also notice the presence of a barely 538 539 expressed forebulge in this area (Fig. 13a, c).

The dichotomy between the northern and southern parts is especially obvious on a N-S 540 transect (ZZ' in Fig. 13d). A shallow southern sub-basin with a gentle northwards dip (< 541 c. 250 < 5 > m deep) is identified, as well as a northern deeper basin with steep borders 542 (c. 600 < 5 > m deep). The limit between the northern and southern parts appears relatively 543 close to the MT/FT boundary (Fig. 13d), suggesting a significant role for lithospheric 544 boundaries in the differential flexuration of the SFB. This N-S differentiation is found not 545 only in the foreland, but also within the allochthon itself as its simulated elevation is not 546 continuous along its front (Fig. 13a). Two areas of important elevations (>1200<5>m) can be 547 observed on both the northern and southern sides of the GA recess. This positive relief could 548 have contributed as a significant source of terrigenous material, then being deposited in the 549 550 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 present in the central SFB and could physically have partly separated these two parts of thebasin.

557 5.a. Evidence for a foreland basin

The convex 'lozenge shape' (sensu Miall, 2010) of the isopach map (Fig. 8) and the 558 559 westwards-thickening pattern of the sedimentary record are in agreement with the common asymmetric geometry of foreland basins (Fig. 8; DeCelles & Giles, 1996; Miall, 2010). 560 Additionally, the observed high-rate subsidence values (c. $100-500 < 5 > m Ma^{-1}$) agree with 561 foreland basin dynamics, even if these values are greater in magnitude than values generally 562 given in the literature for similar contexts (e.g. Xie & Heller, 2009). This difference in 563 magnitude is interpreted by considering that estimations from backstripping analyses are 564 generally proposed for continuous sedimentary series spanning several millions years, if not 565 several tenth of millions years (e.g. Xie & Heller, 2009). Over such long time intervals, the 566 subsidence rate values are less accurate. The high resolution of the timeframe used for the 567 SFB mirrors short-acting structural events in the basin. Similar 'higher than average' values 568 for subsidence rates have been calculated by Chevalier et al. (2003) and Lachkar et al. (2009) 569 using high-resolution biostratigraphic time-calibrations, and also by Roddaz et al. (2010) 570 with similar magnitude for the Miocene Amazonian Foreland Basin (c. 200–700<5>m Ma⁻¹; 571 Roddaz *et al.* 2010). Moreover, values observed in the SFB ($0.05-0.65 \text{ mm a}^{-1}$) are consistent 572 with yearly deposition rates indicated by Allen & Allen (2005) for foreland basins (0.2-573 $0.5 < 5 > \text{mm a}^{-1}$). Finally, the convex-up shape of the tectonic subsidence curves (Fig. 9f) is 574 diagnostic of foreland basins and corresponds to the progressive flexural response of the 575 lithosphere to the topographic load and/or sedimentary infill of the basin overtime (Angevine, 576 Heller & Paola, 1990; Allen & Allen, 2005; Xie & Heller, 2009). 577

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

The observed spatial heterogeneity of the sedimentary thickness in the SFB (Figs 4, 8) and the much higher tectonic subsidence rate detected in the northern part of the basin (c. $500 < 5 > m Ma^{-1} v. c. 100 < 5 > m Ma^{-1}$ in the southern part; Fig. 9f) are striking and raise the

question of the controlling factor(s) responsible for this phenomenon, especially for such a short interval (c. 1.3 < 5 > Ma).

5.b. Potential underlying mechanisms for observed variations in flexuralsubsidence

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

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.

Regarding points (2) and (3) above, while it is possible to discuss the role of the
sedimentary overload using only field-based data, interpretations of the allochthon
heterogeneity and the basement rheological behaviour require an additional model approach.
We combine these in the following discussion. To that purpose, we used three different
scenarios (Fig. 14) with the same initial setup (Section 4.d; Table 2) except for the *x* and *y*dimensions of the model that are set to 2000<5>km in the *x* direction and 1000<5>km in the *y* direction to avoid border effects.

The first scenario tests the impact of a rheologically heterogeneous basement loaded by 626 627 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 sedimentary records are therefore a direct consequence of the rheological behaviour of the 629 630 basement (Angevine, Heller & Paola, 1990; Watts, 1992; Cardozo & Jordan, 2001; Allen & Allen, 2005; Leever et al. 2006; Fosdick, Graham & Hilley, 2014). Upon the high-rigidity 631 part of the basement (T_{e1}), a wide foreland ($\lambda_1 \approx 250 < 5 > \text{km}$) develops with a well-expressed 632 633 convex shape in map view and a barely expressed forebulge. Upon the low-rigidity parts of the basement (T_{e2}), a narrower foreland ($\lambda_2 \approx 110 < 5 > \text{km}$) is structured with a more 634 pronounced forebulge. This is in agreement with the SFB observations. However, a N-S 635 transect (aa', Fig. 14a) shows that the wider area of the foreland basin is deeper than 636 observed in the field and that only one high-relief area is individualized within the central 637 part of the allochthon. Even if the rigidity does play a role in the development of the flexural 638 639 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; Miall, 2010; Fosdick, Graham & Hilley, 2014), our results indicate that a rheological 641 difference is not enough to control the variations in SFB. 642

The second scenario uses a heterogeneous topographic load exerted by the allochthon upon a homogeneous 'strong' lithosphere ($T_e = 90 < 5 > \text{km}$; Fig. 14b). The heterogeneity in the

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

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

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

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 pattern of the basement terranes, the lithosphere heritage likely controls the flexuration andtherefore the subsidence variations documented for the Early Triassic SFB.

Combining all field data and numerical simulations, a model of the SFB is proposed in 676 Figure 15. The northern part of the basin (section AA') is characterized by a narrow foredeep 677 $(\lambda \approx 250 < 5 > \text{km})$ with a high-rate tectonic subsidence (c. 500 < 5 > m Ma⁻¹) and high 678 sedimentary thickness (up to c. 550 < 5 > m of mostly fine siltstones deposits), which is located 679 upon the 'weak/attenuated' Farmington Terrane and in front of the largest reconstructed part 680 of the GA. The postulated wedge-top and forebulge are located above the 'strong' Archean 681 682 lithospheres, that is, the GCB and WT, respectively. The southern part of the SFB exhibits a large foredeep ($\lambda \approx 500 < 5 > \text{km}$, section *BB*') with a relatively low-rate tectonic subsidence 683 (c. $100 < 5 > m \text{ Ma}^{-1}$) and a reduced sedimentary thickness (up to c. 250 < 5 > m of mixed 684 limestones and coarse clastic deposits). This part of the SFB is emplaced upon the 'strong' 685 lithospheres of the Palaeoproterozoic MT and YT, in front of the thinnest reconstructed part 686 687 of the GA. The southern SFB also shows a reduced postulated wedge-top to the west and a barely expressed forebulge to the east. These spatial variations in flexural subsidence and 688 689 their good agreement with limits of the terranes composing the SFB basement are also evident along a N-S transect (section CC'). The spatial separation between the shallow and 690 gently dipping southern part of the SFB and the deep and steep northern part is obvious. This 691 separation is located close to the boundary between MT and FT. 692

693 6. Conclusion

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

Using high-resolution temporal data, the subsidence analyses help to identify the main
 controlling factors at the origin of the spatial variations of the Early Triassic sedimentary
 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 of the contrasted lithospheric strength of the terranes ('weak' v. 'strong' lithospheres) 704 composing the basement of the basin, and the spatial heterogeneity of the Golconda 705 Allochthon (with the presence of a lateral ramp within the belt), best explain a differential 706 707 flexural response of the SFB basement to the emplacement of the allochthon. Such a differential flexural response ultimately controls the overall geometry of the basin through 708 spatially heterogeneous tectonic subsidence rates: c. $100 < 5 > m Ma^{-1}$ in a wide southern part 709 upon a 'strong' lithosphere loaded by a recessed and thin (in map-view) front belt, v. 710 c. $500 < 5 > m \text{ Ma}^{-1}$ in a narrower northern part upon a 'weak/attenuated' lithosphere loaded by 711 a larger front belt. Although field data highlight the potential role of the rheological 712 713 behaviour of the basement based on observed differential subsidence rates, the numerical model approach suggests a combined effect of the latter and of the spatial heterogeneity of 714 715 the allochthon.

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

721 Acknowledgements

We particularly thank the late Professor W.R. Dickinson for constructive discussions. We also thank
Hugo Bucher and Romain Jattiot (Zürich) for discussion on Nevada outcrops. This work is a

rugo Bucher and Roman satisf (Zurien) for discussion on Nevada outcrops. This work is a
 contribution to the ANR project AFTER (ANR-13-JS06-0001-01). The study was also supported by
 ENGIE.

726 Supplementary material

- 727 To view supplementary material for this article, please visit
- 728 https://doi.org/10.1017/S0016756817000164.

729 References

730

731 ALLEN, P. A. & ALLEN, J. R. 2005. Basin Analysis: Principles and Applications. Oxford:

- 732 Blackwell Science Publishing.
- ANGEVINE, C. L., HELLER, P. L. & PAOLA, C. 1990. *Quantitative Sedimentary Basin Modeling*. American Association of Petroleum Geologists.
- ARTEMIEVA, I. M. & MOONEY, W. D. 2002. On the relations between cratonic lithosphere
 thickness, plate motions, and basal drag. *Tectonophysics* 358(1–4), 211–31.
- BANKEY, V. A., CUEVAS, D., DANIELS, A. A., FINN, C. A., HERNANDEZ, I., HILL, P., KUCKS,
 R., MILES, W., PILKINGTON, M., ROBERTS, C., ROEST, W., RYSTROM, V., SHEARER,
- S., SNYDER, S., SWEENEY, R., VELEZ, J., PHILLIPS, J. D. & RAVAT, D. 2002. Digital
 data grids for the magnetic anomaly map of North America. Open-File Report 02-414:
 USGS.
- 742 BEAUMONT, C. 1981. Foreland basins. *Geophysical Journal International* **65**(2), 291–329.
- BLAKEY, R. C. 2008. Pennsylvanian-Jurassic sedimentary basins of the Colorado Plateau and
 Southern Rocky Mountains. *Sedimentary Basins of the World*, pp. 245–96.
- 745 Netherlands: Elsevier.
- BOND, G. C., CHRISTIE-BLICK, N., KOMINZ, M. A. & DEVLIN, W. J. 1985. An early Cambrian
 rift to post-rift transition in the Cordillera of western North America. *Nature* 315,
 742–46.
- 749 BRAYARD, A., BYLUND, K. G., JENKS, J. F., STEPHEN, D. A., OLIVIER, N., ESCARGUEL, G.,
 750 FARA, E. & VENNIN, E. 2013. Smithian ammonoid faunas from Utah: implications for
 751 Early Triassic biostratigraphy, correlation and basinal paleogeography. *Swiss Journal*752 of *Palaeontology* 132(2), 141–219.
- BRAYARD, A., MEIER, M., ESCARGUEL, G., FARA, E., NUETZEL, A., OLIVIER, N., BYLUND, K.
 G., JENKS, J. F., STEPHEN, D. A., HAUTMANN, M., VENNIN, E. & BUCHER, H. 2015.
- Early Triassic 'Gulliver' gastropods; spatio-temporal distribution and significance for
 biotic recovery after the end-Permian mass extinction. *Earth-Science Reviews* 146,
 31–64.
- BRÜHWILER, T., BUCHER, H., BRAYARD, A. & GOUDEMAND, N. 2010. High-resolution
 biochronology and diversity dynamics of the Early Triassic ammonoid recovery; the
 Smithian faunas of the northern Indian margin. *Palaeogeography, Palaeoclimatology, Palaeoecology* 297(2), 491–501.
- BRYANT, B. & NICHOLS, D. 1988. Late Mesozoic and early Tertiary reactivation of an ancient
 crustal boundary along the Uinta trend and its interaction with the Sevier orogenic

- belt. *Geological Society of America Memoirs* **171**, 411–30.
- BUCHER, H. 1988. A new Middle Anisian (Middle Triassic) ammonoid zone from
 northwestern Nevada (USA). *Eclogae Geologicae Helvetiae* 81(3), 723–62.
- BURCHFIEL, B. & DAVIS, G. A. 1975. Nature and controls of Cordilleran orogenesis, western
 United States: Extensions of an earlier synthesis. *American Journal of Science* 275,
 363–96.
- BURCHFIEL, B. & ROYDEN, L. 1991. Antler orogeny: A Mediterranean-type orogeny.
 Geology 19(1), 66–9.
- BURGESS, S. D., BOWRING, S. & SHEN, S.-Z. 2014. High-precision timeline for Earth's most
 severe extinction. *Proceedings of the National Academy of Sciences of the United States of America* 111(9), 3316–21.
- BUROV, E. B. & DIAMENT, M. 1995. The effective elastic thickness (Te) of continental
 lithosphere: What does it really mean? *Journal of Geophysical Research: Solid Earth* **100**(B3), 3905–27.
- CAGNARD, F., BARBEY, P. & GAPAIS, D. 2011. Transition between "Archaean-type" and
 "modern-type" tectonics: insights from the Finnish Lapland Granulite Belt. *Precambrian Research* 187(1), 127–42.
- CARDOZO, N. & JORDAN, T. 2001. Causes of spatially variable tectonic subsidence in the
 Miocene Bermejo Foreland Basin, Argentina. *Basin Research* 13(3), 335–57.
- CHAMBERLAIN, R. L. 1980. Structure and stratigraphy of the Rex Peak Quadrangle, Rich
 County, Utah. *Geology Studies* 27, Part 3, 44–54.
- CHARDON, D., GAPAIS, D. & CAGNARD, F. 2009. Flow of ultra-hot orogens: a view from the
 Precambrian, clues for the Phanerozoic. *Tectonophysics* 477(3), 105–18.
- CHEVALIER, F., GUIRAUD, M., GARCIA, J. P., DOMMERGUES, J. L., QUESNE, D., ALLEMAND,
 P. & DUMONT, T. 2003. Calculating the long-term displacement rates of a normal
 fault from the high-resolution stratigraphic record (early Tethyan rifting, French
- 790 Alps). *Terra Nova* **15**(6), 410–16.
- CLARK, D. L. 1957. Marine Triassic stratigraphy in eastern Great Basin. *AAPG Bulletin*41(10), 2192–222.
- 793 CONEY, P. J. 1987. The regional tectonic setting and possible causes of Cenozoic extension in
- the North American Cordillera. *Geological Society Special Publications* **28**, 177–86.
- 795 CONSTENIUS, K. N., CLARK, D. L., KING, J. K. & EHLER, J. B. 2011. Interim geologic map of

- the Provo 30' x 60' quadrangle, Utah, Wasatch, and Salt Lake Counties, Utah. In *Utah Geological Survey Open-File Report 586DM* (ed. U. G. Survey).
- DECELLES, P. G. & COOGAN, J. C. 2006. Regional structure and kinematic history of the
 Sevier fold-and-thrust belt, central Utah. *Geological Society of America Bulletin*118(7–8), 841–64.
- Becelles, P. G. & Giles, K. A. 1996. Foreland basin systems. *Basin Research* 8(2), 105–
 23.
- DICKERSON, P. W. 2003. Intraplate mountain building in response to continent-continent
 collision; the ancestral Rocky Mountains (North America) and inferences drawn from
 the Tien Shan (Central Asia). *Tectonophysics* 365(1–4), 129–42.
- BOG DICKINSON, W. R. 2002. The Basin and Range Province as a composite extensional domain.
 International Geology Review 44(1), 1–38.
- BO8 DICKINSON, W. R. 2004. Evolution of the North American cordillera. *Annual Review of Earth* and Planetary Sciences 32, 13–45.
- BIO DICKINSON, W. R. 2006. Geotectonic evolution of the Great Basin. *Geosphere* **2**(7), 353–68.

DICKINSON, W. R. 2013. Phanerozoic palinspastic reconstructions of Great Basin
geotectonics (Nevada-Utah, USA). *Geosphere* 9(5), 1384–96.

- B13 DOELLING, H. H. 1980. *Geology and Mineral Resources of Box Elder County, Utah.* Salt
 B14 Lake City, UT: Utah Geological and Mineral Survey.
- EARDLEY, A. J. 1939. Structure of the Wasatch-Great Basin region. *Geological Society of America Bulletin* 50(8), 1277–310.
- EMBRY, A. F. 1997. Global sequence boundaries of the Triassic and their identification in the
 Western Canada sedimentary basin. *Bulletin of Canadian Petroleum Geology* 45(4),
 415–33.
- 820 FAN, M., DECELLES, P. G., GEHRELS, G. E., DETTMAN, D. L., QUADE, J. & PEYTON, S. L.
- 2011. Sedimentology, detrital zircon geochronology, and stable isotope geochemistry
 of the lower Eocene strata in the Wind River Basin, central Wyoming. *Geological Society of America Bulletin* 123(5–6), 979–96.
- FLETCHER, R. C. & HALLET, B. 1983. Unstable extension of the lithosphere: A mechanical
 model for basin-and-range structure. *Journal of Geophysical Research: Solid Earth*826 88(B9), 7457–66.
- 827 FOSDICK, J. C., GRAHAM, S. A. & HILLEY, G. E. 2014. Influence of attenuated lithosphere

- and sediment loading on flexure of the deep-water Magallanes retroarc foreland basin,
 Southern Andes. *Tectonics* 33(12), 2505–25.
- 830 FOSTER, D. A., MUELLER, P. A., MOGK, D. W., WOODEN, J. L. & VOGL, J. J. 2006.
- Proterozoic evolution of the western margin of the Wyoming craton: implications for
 the tectonic and magmatic evolution of the northern Rocky Mountains. *Canadian Journal of Earth Sciences* 43(10), 1601–19.
- GABRIELSE, H., SNYDER, W. S. & STEWART, J. H. 1983. Sonoma orogeny and Permian to
 Triassic tectonism in western North America. *Geology* 11(8), 484–86.
- 836 GALFETTI, T., BUCHER, H., OVTCHAROVA, M., SCHALTEGGER, U., BRAYARD, A.,
- 837 BRÜHWILER, T., GOUDEMAND, N., WEISSERT, H., HOCHULI, P. A., CORDEY, F. &
- 838 GUODUN, K. 2007. Timing of the Early Triassic carbon cycle perturbations inferred
- from new U-Pb ages and ammonoid biochronozones. *Earth and Planetary Science Letters* 258(3–4), 593–604.
- GANS, P. & BOHRSON, W. 1998. Suppression of volcanism during rapid extension in the
 Basin and Range Province, United States. *Science* 279(5347), 66–68.
- GESLIN, J. K. 1998. Distal ancestral Rocky Mountains tectonism: Evolution of the
 Pennsylvanian-Permian Oquirrh–Wood River basin, southern Idaho. *Geological Society of America Bulletin* 110(5), 644–63.
- GILBERT, H., VELASCO, A. A. & ZANDT, G. 2007. Preservation of Proterozoic terrane
 boundaries within the Colorado Plateau and implications for its tectonic evolution. *Earth and Planetary Science Letters* 258(1–2), 237–48.
- GOODSPEED, T. H. & LUCAS, S. G. 2007. Stratigraphy, sedimentology, and sequence
 stratigraphy of the Lower Triassic Sinbad Formation, San Rafael Swell, Utah. *Bulletin*
- *New Mexico Museum of Natural History and Science* **40**, 91–101.
- 852 GROSHONG JR, R. H. 2006. *3-D Structural Geology*. Springer.
- HAQ, B. U., HARDENBOL, J. & VAIL, P. R. 1988. Mesozoic and Cenozoic chronostratigraphy
 and cycles of sea-level change. In:, pp. 72–108. Society of Economic Paleontologists
 and Mineralogists, Special Publication no. 42.
- 856 HECKERT, A. B., CHURE, D. J., VORIS, J. T., HARRISON, A. A. & THOMSON, T. J. 2015.
- 857 Stratigraphy, correlation and age of the Moenkopi Formation in the vicinity of
- 858 Dinosaur National Monument, Eastern Uinta Basin, Utah and Colorado, USA. In
- 859 *Geology of Utah's Uinta Basin and Uinta Mountains* (eds M. D. Vanden Berg, R.

- 860 Ressetar and L. P. Birgenheier), pp. 1–12. Utah Geological Association Publication.
- 861 HOFMANN, R., HAUTMANN, M., BRAYARD, A., NUETZEL, A., BYLUND, K. G., JENKS, J. F.,
- VENNIN, E., OLIVIER, N. & BUCHER, H. 2014. Recovery of benthic marine
 communities from the end-Permian mass extinction at the low latitudes of eastern
 Panthalassa. *Palaeontology* 57(3), 547–89.
- INGERSOLL, R. V. 2008. Subduction-related sedimentary basins of the USA Cordillera. In:
 Sedimentary Basins of the World, pp. 395–428. Amsterdam, Netherlands: Elsevier.
- JATTIOT, R., BUCHER, H., BRAYARD, A., BROSSE, M., JENKS, J. & BYLUND, K. G. (in press).
 Smithian ammonoid faunas from northeastern Nevada: implications for Early Triassic
 biostratigraphy and correlation within the western USA basin. *Palaeontographica Abteilung A*.
- JATTIOT, R., BUCHER, H., BRAYARD, A., MONNET, C., JENKS, J. F. & HAUTMANN, M. 2015.
 Revision of the genus *Anasibirites* Mojsisovics (Ammonoidea): an iconic and
 cosmopolitan taxon of the late Smithian (Early Triassic) extinction. *Papers in Palaeontology* 2(1), 155–188.
- KLUTH, C. F. & CONEY, P. J. 1981. Plate tectonics of the ancestral Rocky Mountains. *Geology* 9(1), 10–15.
- KREEMER, C. & HAMMOND, W. C. 2007. Geodetic constraints on areal changes in the
 Pacific–North America plate boundary zone: What controls Basin and Range
 extension? *Geology* 35(10), 943–46.
- KUCKS, R. P. 1999. Bouguer gravity anomaly data grid for the conterminous US. US
 Geological Survey Digital Data Series DDS-9: USGS.
- KUMMEL, B. 1954. *Triassic Stratigraphy of Southeastern Idaho and Adjacent Areas*. US
 Government Printing Office.
- KUMMEL, B. 1957. Paleoecology of Lower Triassic formations of southeastern Idaho and
 adjacent areas. *Geological Society of America Memoirs* 67, 437–68.
- LACHENBRUCH, A. H. & MORGAN, P. 1990. Continental extension, magmatism and elevation;
 formal relations and rules of thumb. *Tectonophysics* 174(1), 39–62.
- LACHKAR, N., GUIRAUD, M., EL HARFI, A., DOMMERGUES, J.-L., DERA, G. & DURLET, C.
 2009. Early Jurassic normal faulting in a carbonate extensional basin; characterization
- 890 of tectonically driven platform drowning (High Atlas rift, Morocco). *Journal of the*
- 891 *Geological Society of London* **166**(3), 413–30.

- LAWTON, T. F., BOYER, S. E. & SCHMITT, J. G. 1994. Influence of inherited taper on
 structural variability and conglomerate distribution, Cordilleran fold and thrust belt,
 western United States. *Geology* 22(4), 339–42.
- LE POURHIET, L. & SALEEBY, J. 2013. Lithospheric convective instability could induce creep
 along part of the San Andreas fault. *Geology* 41(9), 999–1002.
- LEEVER, K., MATENCO, L., BERTOTTI, G., CLOETINGH, S. & DRIJKONINGEN, G. 2006. Late
 orogenic vertical movements in the Carpathian Bend Zone–seismic constraints on the
 transition zone from orogen to foredeep. *Basin Research* 18(4), 521–45.

900 LOWRIE, W. 2007. Fundamentals of Geophysics. Cambridge: Cambridge University Press.

- 901 LUCAS, S. G., KRAINER, K. & MILNER, A. R. 2007. The type section and age of the
- 902 Timpoweap Member and stratigraphic nomenclature of the Triassic Moenkopi Group
- 903 in Southwestern Utah. Triassic of the American West. *New Mexico Museum of*
- 904 *Natural History and Science Bulletin* **40**, 109–18.
- LUCAS, S. G. & ORCHARD, M. J. 2007. Triassic lithostratigraphy and biostratigraphy north of
 Currie, Elko County, Nevada. *Bulletin New Mexico Museum of Natural History and Science* 40, 119–26.
- 208 LUND, K., BOX, S. E., HOLM-DENOMA, C. S., SAN JUAN, C. A., BLAKELY, R. J., SALTUS, R.
 209 W., ANDERSON, E. D. & DEWITT, E. H. 2015. Basement domain map of the
 210 Anterprint of the States and Alaska Poster, VA, United States US Coolegies
- 910 conterminous United States and Alaska. Reston, VA, United States: US Geological911 Survey.
- MALAVIEILLE, J. 1993. Late orogenic extension in mountain belts: insights from the Basin
 and Range and the late Paleozoic Variscan belt. *Tectonics* 12(5), 1115–30.
- 914 MANGER, G. E. 1963. Porosity and bulk density of sedimentary rocks. USGPO.
- MARZOLF, J. E. 1993. Palinspastic reconstruction of early Mesozoic sedimentary basins near
 the latitude of Las Vegas; implications for the early Mesozoic Cordilleran cratonal
- 917 margin. Field Trip Guidebook Pacific Section, Society of Economic Paleontologists
 918 and Mineralogists 71, 433–62.
- MCCULLOH, T. H. 1967. Mass properties of sedimentary rocks and gravimetric effects of
 petroleum and natural-gas reservoirs. US Government Printing Office.
- MIALL, A. 2010. *The Geology of Stratigraphic Sequences*. Springer Science & Business
 Media.
- 923 MOREAU, J., LE POURHIET, L., HUUSE, M., GIBBARD, P. L. & GRAPPE, B. 2015. The impact

- 924 of the lithospheric flexure during the Elsterian glacial maximum on post-/proglacial
 925 systems in the southern North Sea area. In QRA 2015 Annual Discussion Meeting.
- MUELLER, P. A., WOODEN, J. L., MOGK, D. W. & FOSTER, D. A. 2011. Paleoproterozoic
 evolution of the Farmington Zone; implications for terrane accretion in southwestern
 Laurentia. *Lithosphere* 3(6), 401–08.
- MUKUL, M. & MITRA, G. 1998. Finite strain and strain variation analysis in the Sheeprock
 Thrust Sheet: an internal thrust sheet in the Provo salient of the Sevier Fold-and-

931Thrust belt, Central Utah. Journal of Structural Geology 20(4), 385–405.

- NELSON, S. T., HART, G. L. & FROST, C. D. 2011. A reassessment of Mojavia and a new
 Cheyenne Belt alignment in the eastern Great Basin. *Geosphere* 7(2), 513–27.
- NEWELL, N. D. & KUMMEL, B. 1942. Lower Eo-Triassic stratigraphy, western Wyoming and
 southeast Idaho. *Geological Society of America Bulletin* 53(6), 937–96.
- NICHOLS, K. M. & SILBERLING, N. J. 1977. Stratigraphy and depositional history of the Star
 Peak Group (Triassic), northwestern Nevada. *Geological Society of America Special Papers* 178, 1–74.
- OLDOW, J. S., BALLY, A. W., AVÉ LALLEMANT, H. & LEEMAN, W. P. 1989. Phanerozoic
 evolution of the North American Cordillera; United States and Canada. *The Geology*of North America, 139–232.
- OLIVIER, N., BRAYARD, A., FARA, E., BYLUND, K. G., JENKS, J. F., VENNIN, E., STEPHEN, D.
 A. & ESCARGUEL, G. 2014. Smithian shoreline migrations and depositional settings in
 Timpoweap Canyon (Early Triassic, Utah, USA). *Geological Magazine* 151(5), 938–
 55.
- 946 OLIVIER, N., BRAYARD, A., VENNIN, E., ESCARGUEL, G., FARA, E., BYLUND, K. G., JENKS, J.
- 947 F., CARAVACA, G. & STEPHEN, D. A. 2016. Evolution of depositional settings in the
 948 Torrey area during the Smithian (Early Triassic, Utah, USA) and their significance for
 949 the biotic recovery. *Geological Journal* 51(4), 600–26.
- PARSONS, T., THOMPSON, G. A. & SLEEP, N. H. 1994. Mantle plume influence on the
 Neogene uplift and extension of the US western Cordillera? *Geology* 22(1), 83–6.
- PAULL, R. A. & PAULL, R. K. 1991. Allochthonous rocks from the western part of the early
 Triassic miogeocline; Hawley Creek area, east-central Idaho. *Contributions to Geology* 28(2), 145–54.
- 955 PAULL, R. A. & PAULL, R. K. 1993. Interpretation of Early Triassic nonmarine-marine

- relations, Utah, USA. *New Mexico Museum of Natural History and Science Bulletin* 3,
 403–09.
- PAULSEN, T. & MARSHAK, S. 1999. Origin of the Uinta recess, Sevier fold-thrust belt, Utah:
 influence of basin architecture on fold-thrust belt geometry. *Tectonophysics* 312(2),
 203–16.
- PETERSON, J. A. 1977. Paleozoic shelf-margins and marginal basins, western Rocky
 Mountains–Great Basin, United States. In: *Rocky Mountain Thrust Belt, Geology and Resources* (eds E. L. Helsey *et al.*), 135–53. Guidebook Wyo. Geol. Assoc. Annu.
 Field Conf. 29.
- POUDJOM DJOMANI, Y. H., O'REILLY, S. Y., GRIFFIN, W. L. & MORGAN, P. 2001. The
 density structure of subcontinental lithosphere through time. *Earth and Planetary Science Letters* 184(3–4), 605–21.
- 968 PURUCKER, M. & WHALER, K. 2007. Crustal magnetism. *Treatise on Geophysics* 5, 195–237.
- 969 RIOUX, R. L., HITE, R. J., DYNI, J. R. & GERE, W. C. 1975. Geologic map of the Upper
- 970 Valley Quadrangle, Caribou County, Idaho. Reston, VA, United States: US
- 971 Geological Survey. RODDAZ, M., HERMOZA, W., MORA, A., BABY, P., PARRA, M.,
- 972 CHRISTOPHOUL, F., BRUSSET, S. & ESPURT, N. 2010. Cenozoic sedimentary evolution
- 973 of the Amazonian foreland basin system. In: *Amazonia, Landscape and Species*

Evolution: A Look into the Past. Hoboken: Blackwell-Wiley, 61–88.

- 975 ROWLEY, P. D., VICE, G. S., MCDONALD, R. E., ANDERSON, J. J., MACHETTE, M. N.,
- 976 MAXWELL, D. J., EKREM, E. B., CUNNINGHAM, C. G., STEVEN, T. A. & WARDLAW, B.
- 977 R. 2005. Interim geologic map of the Beaver 30'x60' Quadrangle, Beaver, Piute, Iron,
 978 and Garfield Counties, Utah. Utah Geological Survey, Open-File Report 454, scale
 979 1:100,000.
- SADLER, R. K. 1981. Structure and stratigraphy of the Little Sheep Creek area, Beaverhead
 County, Montana. United States. Published thesis.
- SCHELLING, D. D., STRICKLAND, D. K., JOHNSON, K. R. & VRONA, J. P. 2007. Structural
 geology of the central Utah thrust belt. *Utah Geological Association Publication* 36,
 1–29.
- SCHWEICKERT, R. A. & LAHREN, M. M. 1987. Continuation of Antler and Sonoma orogenic
 belts to the eastern Sierra Nevada, California, and Late Triassic thrusting in a
 compressional arc. *Geology* 15(3), 270–73.

- 988 SCLATER, J. G. & CHRISTIE, P. A. F. 1980. Continental stretching: an explanation of the post-Mid-Cretaceous subsidence of the central North Sea basin. Journal of Geophysical 989 *Research* **85**(B7), 3711–39. 990
- SHANMUGAM, G. & WALKER, K. R. 1980. Sedimentation, subsidence, and evolution of a 991 992 foredeep basin in the Middle Ordovician, southern Appalachians. American Journal of *Science* **280**(6), 479–96. 993
- 994 SNYDER, W. S. & BRUECKNER, H. K. 1983. Tectonic evolution of the Golconda allochthon, Nevada: problems and perspectives. 995
- SPEED, R. & SILBERLING, N. J. 1989. IGC Field Trip T122: Early Mesozoic tectonics of the 996 Western Great Basin, Nevada. In Early Mesozoic Tectonics of the Western Great 997 998 Basin, Nevada: Battle Mountain to Yerington District, Nevada, July 1–7, 1989, 1.
- SPEED, R. & SLEEP, N. 1982. Antler orogeny and foreland basin: a model. Geological Society 999 1000 of America Bulletin 93(9), 815–28.
- 1001 STECKLER, M. & WATTS, A. 1978. Subsidence of the Atlantic-type continental margin off New York. *Earth and Planetary Science Letters* **41**(1), 1–13. 1002
- 1003 STRICKLAND, A., MILLER, E. L. & WOODEN, J. L. 2011. The timing of Tertiary

- metamorphism and deformation in the Albion-Raft River-Grouse Creek metamorphic 1004 core complex, Utah and Idaho. Journal of Geology 119(2), 185–206. 1005
- TENZER, R., SIRGUEY, P., RATTENBURY, M. & NICOLSON, J. 2011. A digital rock density map 1006 of New Zealand. Computers & Geosciences 37(8), 1181-91. 1007
- THOMAZO, C., VENNIN, E., BRAYARD, A., BOUR, I., MATHIEU, O., ELMEKNASSI, S., OLIVIER, 1008
- N., ESCARGUEL, G., BYLUND, K. & JENKS, J. 2016. A diagenetic control on the Early 1010 Triassic Smithian–Spathian carbon isotopic excursions recorded in the marine settings 1011 of the Thaynes Group (Utah, USA). *Geobiology* 14(3), 220–36.
- TREXLER, J. H. & NITCHMAN, S. P. 1990. Sequence stratigraphy and evolution of the Antler 1012 1013 foreland basin, east-central Nevada. Geology 18(5), 422-25.
- TURNER, G., RASSON, J. & REEVES, C. 2007. Observation and measurement techniques. 1014 Treatise in Geophysics, Geomagnetism 5, 93–146. 1015
- VAN HINTE, J. 1978. Geohistory analysis: application of micropaleontology in exploration 1016 geology. *AAPG Bulletin* **62**(2), 201–22. 1017
- VENNIN, E., OLIVIER, N., BRAYARD, A., BOUR, I., THOMAZO, C., ESCARGUEL, G., FARA, E., 1018 1019 BYLUND, K. G., JENKS, J. F., STEPHEN, D. A. & HOFMANN, R. 2015. Microbial

1020 deposits in the aftermath of the end-Permian mass extinction; a diverging case from the Mineral Mountains (Utah, USA). Sedimentology 62(3), 753–92. 1021 VETZ, N. Q. 2011. Geochronologic and isotopic investigation of the Koipato Formation, 1022 northwestern Great Basin, Nevada: implications for Late Permian-Early Triassic 1023 1024 tectonics along the Western US Cordillera. Boise State University. Published thesis. WALKER, J. D. 1985. Permo-Triassic paleogeography and tectonics of the Southwestern 1025 1026 United States. Cambridge: Massachusetts Institute of Technology. Published thesis. 1027 WARE, D., BUCHER, H., BRAYARD, A., SCHNEEBELI-HERMANN, E. & BRÜHWILER, T. 2015. 1028 High-resolution biochronology and diversity dynamics of the Early Triassic ammonoid recovery: the Dienerian faunas of the Northern Indian Margin. 1029 1030 Palaeogeography, Palaeoclimatology, Palaeoecology 440, 363–73. WATTS, A. 1992. The effective elastic thickness of the lithosphere and the evolution of 1031 1032 foreland basins. Basin Research 4(3-4), 169-78. 1033 WATTS, A. B. 2001. Isostasy and Flexure of the Lithosphere. Cambridge: Cambridge University Press. 1034 1035 WHITMEYER, S. J. & KARLSTROM, K. E. 2007. Tectonic model for the Proterozoic growth of North America. Geosphere 3(4), 220–59. 1036 WILKERSON, M. S., APOTRIA, T. & FARID, T. 2002. Interpreting the geologic map expression 1037 of contractional fault-related fold terminations: lateral/oblique ramps versus 1038 displacement gradients. Journal of Structural Geology 24(4), 593-607. 1039 XIE, X. & HELLER, P. L. 2009. Plate tectonics and basin subsidence history. Geological 1040 Society of America Bulletin 121(1–2), 55–64. 1041 YE, H., ROYDEN, L., BURCHFIEL, C. & SCHUEPBACH, M. 1996. Late Paleozoic deformation of 1042 1043 interior North America: the greater Ancestral Rocky Mountains. AAPG bulletin 80(9), 1397-432. 1044 1045 YONKEE, W. A., DEHLER, C. D., LINK, P. K., BALGORD, E. A., KEELEY, J. A., HAYES, D. S., WELLS, M. L., FANNING, C. M. & JOHNSTON, S. M. 2014. Tectono-stratigraphic 1046 1047 framework of Neoproterozoic to Cambrian strata, west-central US: protracted rifting, glaciation, and evolution of the North American Cordilleran margin. Earth-Science 1048 *Reviews* **136**, 59–95. 1049 YONKEE, W. A. & WEIL, A. B. 2010. Reconstructing the kinematic evolution of curved 1050 1051 mountain belts: Internal strain patterns in the Wyoming salient, Sevier thrust belt,

1052	USA. Geological Society of America Bulletin 122 (1–2), 24–49.			
1053	YONKEE, W. A. & WEIL, A. B. 2015. Tectonic evolution of the Sevier and Laramide belts			
1054	within the North American Cordillera orogenic system. Earth-Science Reviews 150,			
1055	531–93.			
1056	6 ZANDT, G., MYERS, S. C. & WALLACE, T. C. 1995. Crust and mantle structure across th			
1057	Basin and Range-Colorado Plateau boundary at 37 N latitude and implications for			
1058	Cenozoic extensional mechanism. Journal of Geophysical Research: Solid Earth			
1059	100 (B6), 10529–48.			
1060	ZIENKIEWICZ, O. C. & TAYLOR, R. L. 2005. The Finite Element Method for Solid and			
1061	Structural Mechanics. Butterworth-Heinemann.			

1062

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

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

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

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

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

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

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

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

Figure 8. (Colour online) Isopach map of the sedimentary thicknesses recorded for the PTU-Smithian interval, showing marked differences in sedimentary thicknesses between northern 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
- studied interval is also indicated (modified from Dickinson, 2013; see also Fig. 12). The
- position of the wedge-top is based on variations in the sedimentary thicknesses and on
- 1127 geophysical data (Fig. 10).
- Figure 9. (Colour online) Subsidence analysis results obtained for the PTU-Smithian interval
 and early Spathian time using 1D backstripping (Steckler & Watts, 1978; Van Hinte, 1978;
 Allen & Allen, 2005). Locations of sections are given in Figure 3b. Ages for the bottom and
- top boundaries of the Smithian are interpolated from ammonoid biozone durations (after
 Brühwiler *et al.* 2010). Sea-level curve after Haq, Hardenbol & Vail (1988). *Ana.*:
- 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. (c) 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)
- 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 and its surroundings (in mGal; after Kucks, 1999). Notable moderate gravity anomalies are 1142 highlighted by a white contour. SRP: Snake River Plain; FA: Farmington Anomaly; SA: 1143 Southern Anomaly. Black lines represent the interpreted remnants of the main geophysical 1144 accidents, and limits between crustal features. (b) Aeromagnetic anomaly map of the Sonoma 1145 Foreland Basin and its surroundings (in nT; after Bankey et al. 2002). Black lines highlight 1146 1147 areas of contrasted magnetic signatures: SRP: Snake River Plain; SZ: Southern magnetic Zone; CZ: Central magnetic Zone; NEZ: North-Eastern magnetic Zone; NZ: Northern 1148 magnetic Zone. (c) Map of the spatial location of the radiochronological ages (U/Pb ages) 1149 after: (1) Foster et al. 2006; (2) Fan et al. 2011; (3) Mueller et al. 2011; (4) Nelson, Hart & 1150 Frost, 2011; (5) Strickland, Miller & Wooden, 2011). Superimposed red dots indicate 1151 Mesoproterozoic metamorphism episodes (Mueller et al. 2011). (d) Map of basement terranes 1152 of the SFB according to their age and nature, with Archean terranes (pale blue), 1153 Palaeoproterozoic terranes (pale green) and Mesoproterozoic mobile belt (pale red). FT: 1154

- 1155 Farmington Terrane; GCB: Grouse Creek Block; MT: Mojave Terrane; WT: Wyoming
- 1156 Terrane; YT: Yavapai Terrane.

Figure 11. (Colour online) Map of the SFB basement (cf. Fig. 10d) after their heritage and
therefore their rheological behaviour. Archean Grouse Creek Block and Wyoming Terrane,
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 (pale1162 red).

1163 Figure 12. (Colour online) (a) Simplified map showing the position of the Uinta recess (lateral ramp) and Wyoming and Central Utah salients (frontal ramps) of the present-day 1164 Sevier TFB (after Paulsen & Marshak, 1999; Yonkee & Weil, 2010) and reconstructed 1165 Golconda Allochthon front and associated recess (lateral ramp). Sedimentary pattern since 1166 1167 Proterozoic time shows two high accommodation zones separated by a topographic high close to the terrane boundaries (Peterson, 1977, Bryant & Nichols, 1988; Paulsen & Marshak, 1168 1999). Palaeolocation of Permian Oquirrh Basin (e.g. Yonkee & Weil, 2015) and documented 1169 PTU-Smithian conglomerates in the western SFB (e.g. Gabrielse, Snyder & Stewart, 1983; 1170 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 conglomerates found in the area delimited in (a), presumably a product of western relief 1173

1174 dismantlement.

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

1186Figure 14. (Colour online) Numerical models showing the effects of the heterogeneities of the

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 heterogenous allochthon with a c. 100<5>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
- a cornering relief appears on both sides of the recessed area in the allochthon. (c) Scenario
- 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<5>km wide recess (lateral ramp).
- 1196 A much wider convex foreland is formed within the recessed area upon the rigid lithosphere,
- and a cornering relief on both sides of the recess in the allochthon is also visible.

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

1213 series and northern silty sedimentation.

1215Table 1. Estimated tectonic transport values used for palinspastic reconstructions of each

s.
2

Sector		Estimated	References
		tectonic	
		transport (km)	
1	Sevier foreland	0	DeCelles & Coogan, 2006; Schelling et al. 2007;
			Yonkee & Weil, 2010; Yonkee et al. 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 et al. 2007
6	Central Utah salient, south sector	<i>c</i> . 75	DeCelles & Coogan, 2006; Schelling et al. 2007
7	Sevier hinterland, Basin &	<i>c</i> . 80	Yonkee et al. 2014
	Range province		

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 <i>E</i> (GPa)	80	80	80	80
Poisson's ratio, v	0.25	0.25	0.25	0.25
Elastic thickness of 'strong' lithosphere, T_{e1}	90	90	90	90
(km)				
Elastic thickness of 'weak' lithosphere, T_{e2}	30	30	n/a	30
<u>(km)</u>				
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, $\rho_{\rm m}$ (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









































Homogeneous basement & recessed allochthon

q

200 km

Ε



200 km

υ

Heterogeneous basement & recessed allochthon







С