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Paleoproterozoic gold events in the southern West African Craton: review and synopsis

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1 **Gold metallogeny of the Paleoproterozoic part of the southern West**

2 **African Craton: capturing distinct pulses within an orogenic cycle**

3

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23

24 **Abstract**

25

26 The southern part of the West African Craton includes the Baoulé-Mossi Domain, the world's
27 premier Paleoproterozoic gold province (~10,000 metric ton gold endowment). Structural,
28 metamorphic, and geochronological data suggest gold mineralisation occurred during three

29 episodes that span much of the Eoeburnean and Eburnean orogenic cycles. Eoeburnean
30 orogenic and rare skarn-hosted gold deposits formed between ca. 2200 and 2135 Ma during
31 repeated episodes of volcanism, plutonism, and shortening, which thickened the
32 Paleoproterozoic crust. Early Eburnean orogenic and placer gold deposits formed between ca.
33 2110 and 2095 Ma during inversion, metamorphism, and subsequent oblique shortening of
34 intra-orogenic basins filled after ca. 2135 Ma. This episode of mineralisation terminated
35 when the Baoulé-Mossi Domain docked with the Archean Kénéma-Man Domain at ca. 2095
36 Ma. Late Eburnean orogenic and less common intrusion-related gold deposits formed
37 between ca. 2095 and 2060 Ma during strike-slip to oblique-slip tectonics, post-collisional
38 high-K plutonism and crustal reworking across the western and southern Baoulé-Mossi
39 Domain. Eoeburnean gold deposits include ca. 10 % of the gold endowment of the Baoulé-
40 Mossi Domain, whereas the Early Eburnean and Late Eburnean deposits include ca. 50–70%
41 and 20–40%, respectively. Here, we highlight the favourable confluence of accretion-
42 collision tectonics, involving juvenile crust formation as well as protracted magmatic,
43 metamorphic, and deformation histories that resulted in diachronous gold events spread over
44 at least 100 myr throughout the Baoulé-Mossi Domain.

45

46

47 **Keywords:** West African Craton, Paleoproterozoic, gold, Eburnean Orogeny

48

49 **Introduction**

50

51 Tectonic and associated geodynamic processes ultimately control the character, timing, and
52 distribution of ore-forming mineral systems (Robert et al. 2005; Kerrich et al. 2005; Bierlein
53 et al. 2006; McCuaig and Hronsky 2014). Developing an understanding of ore-forming

54 processes at the scale of a craton is commonly hindered by the complexity of the geological
55 record derived from deposit and camp scale studies. Investigations of gold deposits in the
56 Paleoproterozoic Baoulé-Mossi Domain of the southern part of the West African Craton
57 (sWAC) have studied the age and geological setting of gold mineralisation within specific
58 deposits and camps (Wassa, Perrouy et al. 2016; Siguiri, Lebrun et al. 2017; Obuasi,
59 Blenkinsop et al. 1994; Allibone et al. 2002a; Fougrouse et al. 2017; Oliver et al. 2020;
60 Pampe, Salvi et al. 2016a; Sadiola, Masurel et al. 2017a; Kiaka, Fontaine et al. 2017; Mana,
61 Augustin et al. 2017; Loulo, Lawrence et al. 2013a; Allibone et al. 2020). These studies
62 indicate that gold deposits formed diachronously throughout >100 myr of Paleoproterozoic
63 crustal growth within the Baoulé-Mossi Domain (Goldfarb et al. 2017; Le Mignot et al.
64 2017b). Volcanic, volcano-sedimentary, and plutonic rocks of the Baoulé-Mossi Domain
65 were deposited and emplaced during several major episodes of Paleoproterozoic crustal
66 growth between ca. 2300 and 2060 Ma (Abouchami et al. 1990; Boher et al. 1992; Sylvester
67 and Attoh 1992; Baratoux et al. 2011; Grenh lm et al. 2019; M riaud et al. 2020), including
68 the ca. 2200–2135 Ma Eoeburnean phase and the ca. 2135–2060 Ma Eburnean phase
69 (Feybesse et al. 2006; Parra-Avila et al. 2016, 2017). Deciphering the tectonic, stratigraphic,
70 and magmatic evolution of the Baoul -Mossi Domain through space and time has improved
71 our ability to explain the heterogeneous distribution of gold mineralisation in order to target
72 prospective areas more effectively (Groves and Bierlein 2007).

73 In this review, we integrate published data acquired from deposit through craton scales to
74 propose a revised tectono-stratigraphic interpretation of the Baoul -Mossi Domain. We then
75 use this interpretation to place the various gold deposits in a craton-wide temporal and spatial
76 context. Lastly, we compare the geological and metallogenic history of the Paleoproterozoic
77 sWAC (Baoul -Mossi Domain) with that of the Neoproterozoic Superior (Canada) and Yilgarn
78 (Western Australia) cratons.

79 **The geological history of the Baoulé-Mossi Domain**

80 The southern portion of the West African Craton (Fig. 1) comprises the Paleoproterozoic
81 Baoulé-Mossi Domain and the Archean Kénéma-Man Domain (Bonhomme 1962; Boher et
82 al. 1992; Feybesse et al. 2006). This paper focuses
83 on the Paleoproterozoic Baoulé-Mossi Domain, which in- cludes elongate belts of volcanic,
84 volcano-sedimentary, and sedimentary rocks separated by younger sedimentary basins and
85 plutonic domains. The following description of the geol- ogy of the Baoulé-Mossi Domain
86 provides the geological con- text of gold mineralisation during Paleoproterozoic orogenic
87 activity. Geochronological data discussed in this review is provided in Electronic
88 Supplementary Data (ESM 1).

89

90 **Juvenile mafic volcanism (ca. 2350–2200 Ma)**

91 Tholeiitic basalts are the oldest supracrustal rocks in the Baoulé-Mossi Domain and dominate
92 the basal part of the lithostratigraphic record in each of the older belts of volcanic and
93 volcano-sedimentary rocks (Fig. 1) across the region (Houndé-Boromo belt, Baratoux et al.
94 2011; Nangodi belt, Sylvester and Attoh 1992; Sefwi belt, Hirdes et al. 1993; Ashanti belt,
95 Dampare et al. 2008; Mako belt, Dioh 1986; Diallo 1994; Dia et al. 1997; Abouchami et al.
96 1990; Labou et al. 2020). The timing of the onset of tholeiitic volcanism remains poorly
97 constrained, but the presence of inherited zir- cons ca. 2350–2300 Ma old throughout the
98 Baoulé-Mossi Domain, irrespective of their proximity to identified Archean crustal material,
99 suggest the existence of mafic crust by ca. 2350 Ma. These ages include the following: (i) the
100 2386 ± 42 to 2258 ± 19 Ma inherited zircon cores from the 2195 ± 4 Ma Ifanteyire granitic
101 gneiss in the Bolé-Navrongo belt in NW Ghana (Siegfried et al. 2009); (ii) the 2295 ± 19 to
102 2285 ± 11 Ma xenocrystic zircons from the 2253 ± 15 Ma Kel Enguef migmatitic orthogneiss
103 in the Ouadalan-Gorouol belt in north- ern Burkina Faso (Tshibubudze et al. 2013); (iii) the

104 2312 ± 17 Ma inherited zircon cores from the 2154 ± 1 Ma Dabakala tonalite (Gasquet et al.
105 2003) in central Côte d'Ivoire; and (iv) the 2308 ± 24 to 2219 ± 24 Ma xenocrystic zircons
106 from the 2176 ± 8 Ma Siou granodiorite in the Houndé belt in Burkina Faso (Augustin et al.
107 2017). Model ages obtained from Lu-Hf analyses on magmatic zircons (Parra-Avila et al.
108 2016; Petersson et al. 2016) and Sm-Nd whole rock analyses (Abouchami et al. 1990; Boher
109 et al. 1992; Doumbia et al. 1998) overlap with this range of xenocrystic zircon ages. Detrital
110 zircon grains in younger sedimentary rocks gradually increase in abundance after ca. 2250
111 Ma, with the most significant increase occurring post ca. 2200 Ma (Grenhölml et al. 2019).
112 The minimum age of tholeiitic volcanism is indirectly constrained by the transition to
113 intermediate-felsic volcanism and the widespread onset of diorite-granodiorite-tonalite-
114 trondhjemitic plutonism at ca. 2200 Ma (De Kock et al. 2011; Parra-Avila et al. 2017). The
115 geodynamic context of the juvenile tholeiitic oceanic crust remains uncertain, although
116 plume-related oceanic plateau (Abouchami et al. 1990; Boher et al. 1992; Augustin and
117 Gaboury 2017), island arc (Sylvester and Attoh 1992; Baratoux et al. 2011; Labou et al.
118 2020), back-arc basin (Vidal and Alric 1994), and mid-ocean ridge (Pouclet et al. 2006;
119 Labou et al. 2020) settings have been proposed.

120

121 ***Arc constructional stage, Eoeburnean shortening, and crustal thickening (ca. 2200–2135***
122 ***Ma)***

123 Lithostratigraphic studies of various volcano-sedimentary belts in the Baoulé-Mossi domain
124 indicate calc-alkaline supracrustal rock successions overlie the basal tholeiitic flows, locally
125 with unconformable contacts (Davis et al. 2015). For example, the supracrustal rock
126 succession in the Oumé-Féttékro belt in central Côte d'Ivoire comprises basal mafic to
127 intermediate lavas and tuffs, overlain by felsic lavas, lapilli tuffs, and ignimbrites intercalated
128 with bimodal volcanoclastic, and minor epiclastic rocks, all interpreted to have been deposited

129 in an evolving volcanic arc ([Mortimer 1990, 1992a, 1992b, 2016](#)). The transition from mafic
130 to intermediate volcanism in the Oumé-Féttékro belt occurred between ~2185 and 2160 Ma
131 ([Kouamélan 1996](#)), which is consistent with ages reported for similar rock sequences
132 elsewhere in the sWAC (e.g. northwestern Ghana, [De Kock et al. 2009](#); [Block et al. 2016](#);
133 southwestern Ghana, [Agyei-Duodu et al. 2009](#); [Hirdes et al. 2007](#); [McFarlane 2018](#); eastern
134 Senegal, [Dia et al. 1997](#); [Hirdes and Davis 2002](#); [Dioh et al. 2006](#); [Gueye et al. 2007](#)).
135 Limited geochronology suggests the onset of diorite-granodiorite-tonalite-trondhjemite
136 plutonism occurred as early as ~2250 Ma but became widespread across the sWAC after
137 ~2200 Ma, with peak in magma flux between ~2160 and 2150 Ma ([Agyei-Duodu et al. 2009](#);
138 [De Kock et al. 2011](#); [Tshibubudze et al. 2015](#); [Petersson et al. 2016](#); [Parra-Avila et al. 2017](#),
139 [2018, 2019](#)). However, the age and chemistry of extensive granitoid terranes between the
140 greenstone belts and sedimentary basins remain largely unknown. These calc-alkaline
141 plutonic rocks are not TTGs as per the definition of [Moyen and Martin \(2012\)](#), but were
142 generated by partial melting of a mafic source ([Ganne et al. 2011](#)). They intrude both the
143 large granitoid domains and adjacent belts of volcano-sedimentary rocks throughout the
144 Baoulé-Mossi domain.

145 The onset of deformation in the Baoulé-Mossi domain followed the switch from basic
146 volcanism, regional extension and/or oceanic rifting of uncertain origin to a demonstrably
147 active convergent margin setting with associated intermediate to felsic volcanism. Structural
148 evidence associated with this early deformation, such as layer-parallel foliation and thrusts,
149 was recognized across the Goren ([Hein 2010](#)), Bouroum-Yalago ([Vanin et al. 2004](#)),
150 Bélahouro ([McCuaig et al. 2016](#)), Ouadalan-Gorouol ([Tshibubuze et al. 2015](#)), and Fada
151 N’Gourma ([Naba et al. 2004](#); [Vegas et al. 2008](#); [Fontaine et al. 2017](#)) belts in Burkina Faso,
152 and the Bolé-Bulenga Domain in NW Ghana ([Block et al. 2016](#)). Although variability in the
153 strain regime has been documented across distinct volcano-sedimentary belts and granitoid

154 domains, this early deformation is correlated with crustal thickening (clockwise P - T - t paths),
155 high-grade metamorphism (amphibolite- to granulite-facies), and subsequent partial melting
156 of the crust (Ganne et al. 2014; Block et al. 2015). The style of this early deformation reflects
157 the dominance of horizontal bulk crustal shortening over more local accommodation
158 mechanisms such as ascending magmas into the crust, and/or doming of partially molten
159 crust (Baratoux et al. 2011; Block et al. 2016). The maximum age for the onset of
160 deformation remains poorly constrained but occurred sometime after ~2200 Ma. The
161 minimum age for crustal thickening and partial melting was interpreted from leucosome-
162 hosted monazite grains dated at 2137 ± 8 Ma in NW Ghana (Block et al. 2016). Peak
163 metamorphism in the Mana region in western Burkina Faso has been similarly dated at 2135
164 ± 11 Ma and correlated with metamorphic zircon overgrowths on magmatic zircons from the
165 2176 ± 8 Ma Siou granodiorite (U-Pb on zircon, Augustin et al. 2017). These results are also
166 consistent with HP - LT blueschist-facies series metamorphism ($P > 10$ kbar, $T < 450^\circ\text{C}$) of
167 supracrustal rocks in the Tenkodogo Fada N’Gourma domain in eastern Burkina Faso (Ganne
168 et al. 2011). These early tectono-metamorphic events have been historically referred to as
169 “Tangaeen” (Hein et al. 2010 in Burkina Faso) or “Eoeburnean” (De Kock et al. 2011 in
170 northwest Ghana). The Eoeburnean orogeny culminated with the collision between
171 independent crustal blocks in modern-day northwest Ghana and southwest Burkina Faso at
172 ca. 2135 Ma (Baratoux et al., 2011; Block et al., 2016). This collision was broadly coeval
173 with the short-lived period of reworking of Archean crust between ca. 2140-2130 Ma
174 documented in the Kibi belt and Suhum basin in southeastern Ghana (Leube et al., 1990;
175 Taylor et al., 1992; Petersson et al., 2016, 2018). Some authors associate this phase of early
176 convergence with N-S-directed bulk crustal shortening based on the rarely observed
177 orientations of ENE-striking fabrics and structures in the field (Tshibubudze et al. 2015;
178 Block et al. 2016). These same authors interpreted the later switch to WNW-ESE to NW-SE-

179 directed bulk crustal shortening at ~2120 Ma to reflect a change in boundary conditions
180 applied to the lithosphere, such as a change in the vergence of subduction and/or direction of
181 plate convergence at the onset of the “Eburnean” orogeny. Alternatively, the
182 Paleoproterozoic tectonic evolution of the sWAC may reflect incremental/progressive NW-
183 SE-directed bulk crustal shortening (Milési et al. 1991; 1992; Eisenlohr and Hirdes 1992;
184 Feybesse et al. 2006; Grenhölml et al. 2019), with these apparent structural components
185 reflecting (i) heterogeneous strain distribution and/or (ii) the domainal development of
186 structural fabrics and/or (iii) the rotation of early fabrics and structures during overprinting
187 deformation.

188

189 *Deep water sedimentary series of the Baoulé-Mossi Domain (ca. 2135–2105 Ma)*

190 Thick sedimentary rock series dominated by flysch sequences were deposited in major
191 basins, some of which were > 200 km across, between ~2135 and 2105 Ma. The deposition
192 of voluminous flysch sequences (e.g. Kumasi in Ghana, Sunyani-Comoé in Côte d’Ivoire,
193 Bagoé in Mali, Siguiri in Guinea, Dialé-Daléma in Senegal, Fig. 2a), derived from the
194 erosion of continental crust constrains the minimum age of uplift following the
195 aforementioned period(s) of metamorphism and deformation in the Baoulé-Mossi domain.
196 Major shear zones that commonly bound these basins probably initiated as extensional faults
197 during this period. These fundamental structures persist throughout the following history of
198 the craton outlined below, being reactivated several times, including during gold
199 mineralization. Available geochronology data suggest diachronous development of these
200 sedimentary basins, with an apparent younging in maximum depositional ages from SE to
201 NW across the Baoulé-Mossi domain (Fig. 2a). This apparent diachroneity appears further
202 supported by the general migration of a magmatic and tectono-metamorphic fronts across the
203 sWAC from southeast to northwest with a ~5 to 15 myr “delay”, as first proposed by Hirdes

204 [et al. \(1996, 2002\)](#) and recently by [Parra-Avila et al. \(2017\)](#). The following description of the
205 basin lithologies and their ages is largely a synthesis of data presented in [Davis et al. \(2015\)](#)
206 and [Grenhölml et al. \(2019\)](#), which include comprehensive reviews of the ages of detrital
207 zircon grains from clastic and volcano-sedimentary series in the Baoulé-Mossi domain.
208 Flysch-type sedimentary sequences that include sandstones, siltstones, and argillites with
209 minor volcanoclastic rocks, chemical sedimentary rocks and volcanic rocks, were deposited in
210 large scale basins across the Baoulé-Mossi domain ([Milési et al. 1989](#); [Feybesse et al. 2006](#)),
211 and together with the older volcano-sedimentary belts define the first-order architecture of the
212 craton. These thick deep-water rock sequences have a similar overall stratigraphy at the scale
213 of the sWAC. In NW Ghana, the Maluwe series was deposited between ~2135 and 2125 Ma
214 ([De Kock et al. 2009, 2011](#)). In SW Ghana, the Kumasi and Sunyani-Comoé series were
215 deposited between ~2135 and 2110 Ma ([Davis et al. 1994](#); [Oberthür et al. 1998](#); [Hirdes et al.](#)
216 [2007](#); [Vidal et al. 2009](#); [Agyei-Duodu et al. 2009](#); [Adadey et al. 2009](#); [Grenhölml et al. 2019](#)).
217 In contrast, the Bagoé series in Mali ([Liégois et al. 1991](#); [Wane et al. 2018](#)), the Siguiri series
218 in Guinea ([Lebrun et al. 2016](#)), and Kofi-Dialé-Daléma series in the Kédougou-Kéniéba inlier
219 ([Hein et al. 2015](#); [Lambert-Smith et al. 2016](#); [Masurel et al. 2017c](#); [Allibone et al. 2019](#)) were
220 deposited between ~2110 and 2095, ~2115 and 2095 Ma, and ~2120 and 2095 Ma,
221 respectively. The restricted presence of impure carbonate rocks in the Siguiri, Kofi, and
222 Dialé-Daléma series ([Milési et al. 1989](#); [Ledru et al. 1991](#); [Lebrun et al. 2016](#); [Lambert-Smith](#)
223 [et al. 2016](#)) together with the similar maximum and minimum depositional ages suggest that
224 these carbonate formations belong to the same regional-scale sedimentary basin and may
225 reflect a hinterland-foreland setting relationship between eastern and western sWAC as first
226 proposed by [Grenhölml et al. \(2019\)](#).
227 No consensus yet exists as to the broader tectonic setting in which these basins formed across
228 the Baoulé-Mossi Domain. Some authors favour a syn-orogenic foreland basin setting

229 (Feybesse et al. 2006; Block et al. 2016), whereas others fa-
230 vour back-arc and fore-arc basin
231 settings controlled by exten- sional detachments (Perrouy et al. 2012) or transtensional
232 deformation during lateral escape of Paleoproterozoic crust against Archean indenters
233 (Grenhölml et al. 2019). The onset of basin formation in the eastern Baoulé-Mossi Domain
234 broadly coincided with the cessation of the Eoeburnean orog- eny at ca. 2140 Ma. However,
235 this temporal relationship does not support or preclude any of the alternative tectonic settings
236 for basin formation.

237 *Shallow water to subaerial sedimentary series of the eastern Baoulé-Mossi Domain (ca.*
238 *2130–2120 Ma)*

239 Coarser grained clastic rock sequences than those present in the aforementioned large-scale
240 basins were locally deposited in northwest Ghana between ca. 2130 and 2120 Ma (Grenhölml
241 et al. 2019). These rocks, referred to as the Bui (Kiessling et al. 1997) and Chako (Block et
242 al. 2016) series, largely consist of sandstones and conglomerates, and have been interpreted
243 as having been deposited predominantly in shallow water to subaerial settings (Grenhölml et
244 al. 2019). The deposition of these series partly overlapped with deposi- tion of the more
245 extensive deep water series in larger scale basins throughout the Baoulé-Mossi Domain. The
246 Chako se- ries was deposited on older Eoeburnean-aged rocks, whereas Zitsmann et al.
247 (1997) indicated that the Bui series overlies the Maluwe deep water series with a conformable
248 to paraconformable contact (Fig. 2a). The Bui and Chako series are interpreted as molasse,
249 which was deposited in local depocentres following the collision between Eoeburnean- aged
250 terranes of northwest Ghana and southwest Burkina Faso and the block of Archean material
251 now confined to the lower crust of southeast Ghana and Togo at ca. 2135 Ma (De Kock et al.
252 2011; Block et al. 2016).

253

254

255 *Deformation and metamorphism of the deep water sedimentary series during the onset of*
256 *the Eburnean orogeny (ca. 2110–2095 Ma)*

257 Deep water sedimentary series and host basins throughout the Baoulé-Mossi Domain were
258 deformed and metamorphosed under lower to upper greenschist-facies conditions between ca.
259 2110 and 2095 Ma, during the early stages of the Eburnean orogeny (Perrouy et al. 2012;
260 Lebrun et al. 2016; Oliver et al. 2020; Allibone et al. 2020). Deformed rocks of the Kumasi
261 series were intruded by calc-alkaline biotite-gran- ites between ca. 2110 and 2105 Ma, and
262 these intrusive rocks were affected by Eburnean deformation and metamorphism indicating
263 both occurred after ca. 2110–2105 Ma (Hirdes et al. 1992, 1993; Hirdes and Nunoo 1994;
264 Oberthür et al. 1998; Adadey et al. 2009; Oliver et al. 2020). Deformed rocks of the Kofi
265 series in the Kédougou-Kéniéba Inlier were intrud- ed by calc-alkaline diorite and quartz-
266 feldspar-porphyry at ca. 2105 Ma (Hein et al. 2015; Masurel et al. 2017c). Inversion of the
267 Siguiri basin had occurred by ca. 2095 Ma when deformed sedimentary rock sequences were
268 unconformably overlain by volcanic rocks of the Bandamian series (Allibone et al. 2020;
269 Mériaud et al. 2020).

270 Multistage shortening and deformation of these sedi- mentary series between ca. 2110 and
271 2095 Ma resulted in the development of 1–10-km-scale isoclinal folds, associ- ated axial-
272 planar cleavage, major intra-basin reverse shear zones, and reactivation of the basin margin
273 faults as major reverse shear zones (Hirdes et al. 1992; Allibone et al. 2002a, 2002b; Oliver et
274 al. 2020). Limited mapping indi- cates that the deep water series now comprise kilometre-
275 scale isoclinally folded slices imbricated along intra-basin faults that likely evolved from
276 major isoclinal folds as shortening progressed. No intact, complete stratigraphic section has
277 been recognised to date, precluding constraints on the original thickness of any deep water
278 series. These folds and the major intra-basin and basin margin faults are the principal

279 structures, which define the NE- to NNE- striking regional scale structural architecture
280 throughout much of the Baoulé-Mossi Domain, obvious in geological maps of the craton
281 (Fig. 1). Syn-metamorphic quartz-carbonate veining occurred throughout the history of
282 basin inversion (Fougerouse et al. 2017; Oliver et al. 2020).

283 Inversion of the Akyem, Kumasi, and Sunyani-Comoé basins in the southeastern part of the
284 Baoulé-Mossi Domain (Fig. 1) may have coincided with an inferred oceanic arc-arc collision
285 between terranes in modern-day southern Ghana and central Ghana–eastern Côte d’Ivoire at
286 ca. 2100 Ma, with the suture located along what is now the northwest margin of the Sefwi
287 belt (McFarlane et al. 2019a). Inversion of the Bagoé, Siguiri, Kofī, and Dialé-Daléma basins
288 in the northwest part of the Baoulé-Mossi Domain occurred before docking of the Baoulé-
289 Mossi and Archean Kénéma-Man domains at ca. 2095 Ma (Lahondère et al. 2002; Eglinger et
290 al. 2017; Thébaud et al. 2020; Mériaud et al. 2020). Field structural relationships imply
291 docking of the Archean and Paleoproterozoic domains was associated with region-wide
292 dextral reactivation of major shear zones such as the Sassandra fault, Banifing shear zone,
293 and Seikorole shear zone (Fig. 1), whereas basin inversion occurred during earlier E-W
294 directed shortening. Deformation of the deep water series in the northwest part of the Baoulé-
295 Mossi Domain may therefore have been a prelude to docking of the Paleoproterozoic and
296 Archean domains.

297

298 ***Shallow water to subaerial sedimentary series of the Baoulé-Mossi Domain (ca. 2110–2095***
299 ***Ma)***

300 Shallow water to subaerial sedimentary series were locally deposited in restricted basins
301 across the Baoulé-Mossi Domain between ca. 2110 and 2095 Ma (Fig. 2a). Sedimentary
302 rocks of these series largely comprise conglomerates and cross-stratified to planar-bedded
303 sandstones, with subordinate siltstones and rhyolitic volcanic rocks also occurring locally in

304 some series. Conglomerates are typically polymict and include clasts of granite, Eoeburnean
305 supracrustal rocks, vein, and meta- morphic quartz, and locally Archean detritus (Milési et al.
306 1989; Feybesse and Milési 1994; Parra-Avila et al. 2015; Lebrun et al. 2016). The Tarkwa
307 (Sestini 1973; Pigois et al. 2003), Safane-Téhini (Simeon et al. 1995; Bossière et al. 1996),
308 and Kintinian (Lebrun et al. 2016) series are con- sistently younger in age than the deep water
309 sedimentary rock sequences described above (Davis et al. 2015; Grenhölml et al. 2019). The
310 Tarkwa series of SW Ghana was deposited between ca. 2110 and 2095 Ma immediately east
311 of the Ashanti shear zone (Davis et al. 1994; Hirdes and Nunoo 1994; Pigois et al. 2003;
312 Oberthür et al. 1998; Adadey et al. 2009; Perrouty et al. 2012). Deposition of the Tarkwa
313 series was coeval with intrusion of ca. 2100 Ma gabbro sills into the clastic sedimentary rocks
314 (Adadey et al. 2009; Perrouty et al. 2012; White et al. 2015). These shallow water to
315 subaerial series have been interpreted as molasse deposited in short-lived basins be- cause
316 their deposition ages overlap with the crystallisation ages of crosscutting syn-kinematic
317 granites (Tunks et al. 2004; Goldfarb et al. 2017). Their internal structural char- acter implies
318 they were deposited during incremental NW- SE-directed bulk crustal shortening towards the
319 culmina- tion of the Eburnean orogeny (Milési et al. 1992; Oberthür et al. 1998; Feybesse et
320 al. 2006; Baratoux et al. 2011; Masurel et al. 2017c; Wane et al. 2018).

321

322 ***Docking of the Archean Kénéma-Man and Paleoproterozoic Baoulé-Mossi Domains (ca.***
323 ***2095 Ma)***

324 High-K calc-alkaline biotite-granite to leucogranite (ca. 2100– 2075 Ma) stitch the Archean
325 rocks of the Kénéma-Man Domain to deformed and metamorphosed Paleoproterozoic
326 sedimentary rocks of the Siguri Series (Lebrun et al. 2016; Eglinger et al. 2017). These
327 relationships indicate docking of the Archean and Paleoproterozoic rocks occurred at ca.
328 $2095 \pm 5\text{--}10$ Ma. Docking of the two domains appears to have coincided with dextral

329 reactivation of the regional scale Sassandra, Banifing, and Féréké shear zones across central
330 and western parts of the Baoulé-Mossi Domain (Fig. 1). Associated deformation may also
331 have resulted in the oroclinal bending of the Siguri Series and Seikorole shear zone, such that
332 southern parts of the series and shear zone now strike east-west parallel to the margin of the
333 Kénéma- Man Domain, whereas the general structural grain of rocks to the north and east
334 typically strikes north to NNE (Fig. 1). The direction of convergence between the two
335 domains re- mains unclear, although the general regional-scale structural architecture implies
336 the Kénéma-Man Domain approached the Baoulé-Mossi Domain from either the southwest
337 or west, in- denting the southwest margin of the Baoulé-Mossi Domain.

338

339 ***Late Eburnean strike-slip to oblique-slip tectonics and crustal reworking following arc-***
340 ***continent collision (ca. 2095–2060 Ma)***

341 Two or more diachronous phases of either strike-slip and/or oblique-slip displacement
342 occurred along pre-existing faults and shear zones throughout southern and western parts of
343 the Baoulé-Mossi Domain between ca. 2095 and 2060 Ma (Ledru et al. 1991; Milési et al.
344 1992; Allibone et al. 2002a, 2002b; McFarlane et al. 2019b; Oliver et al. 2020). Strain during
345 this period was largely localised along major shear zones that form geological domain
346 boundaries with little mod- ification to the pre-existing structural architecture (Fig. 2b). For
347 example, sinistral displacement is recorded on the WNW- to NNW-striking Sassandra shear
348 zone after ca. 2100 Ma (Egal et al. 2002; Lebrun et al. 2017) and along the NNE-striking
349 Banifing and Syama shear zones in Mali (Liégeois et al. 1991; Caby et al. 2000; Ballo et al.
350 2016; Wane et al. 2018), contemporaneous with emplacement of the 2096 ± 7 Ma Massigui
351 granite (Parra-Avila et al. 2017). In the Kédougou-Kéniéba Inlier, the switch from bulk
352 orogen- perpendicular E-W shortening to oblique convergence oc- curred at ca. 2085 Ma,
353 contemporaneously with magmatism in the Falémé area (Masurel et al. 2017b, 2017c;

354 Allibone et al. 2020). In SW Ghana, segments of the Ashanti- Konongo-Prestea, Bibiani, and
355 Kenyase-Yamfo shear zones were reactivated as sinistral-reverse shear zones between ca.
356 2105 and 2085 Ma (Allibone et al. 2002a, 2002b, 2004; Feybesse et al. 2006; Oliver et al.
357 2020). In central Côte d'Ivoire, sinistral displacement along the N'Zi-Brobo shear zone
358 occurred after ca. 2100 Ma (Mortimer 1990; Leake 1992). Relatively minor, but
359 economically significant, N- to NNE-striking primary strike-slip faults that refract across the
360 earlier structural grain also developed during this period (Allibone et al. 2002a).
361 Amphibolite- and granulite-facies rocks were unroofed and juxtaposed against mid-crustal
362 greenschist-facies rocks during transtension in the northwest part of the Sefwi belt and nearby
363 Chiraa domain in the adjacent Sunyani-Comoé basin at 2073 ± 4 Ma (McFarlane et al.
364 2019a), indicating extension occurred throughout the full thickness of the crust locally within
365 the southern and western Baoulé-Mossi Domain during this period. Granulite-facies mineral
366 assemblages developed within ductile shear zones along the northern Kénéma-Man Domain
367 margin at 2071 ± 2 Ma (Feybesse et al. 2000; Lerouge et al. 2004). In contrast, deformation
368 further away from this boundary occurred under greenschist-facies brittle-ductile conditions
369 and retrograde P- T paths (in the Kédougou-Kéniéba inlier, Masurel et al. 2017c; in western
370 and central Côte d'Ivoire, Mortimer 1990; Leake 1992; Kouamélan 1996).
371 Oblique-slip reactivation of shear zones along the northern margin of the Archean Kénéma-
372 Man Domain coincided with the emplacement of voluminous high-K calc-alkaline to
373 shoshonitic granites between ca. 2100 and 2075 Ma (Feybesse and Milési 1994; Egal et al.
374 2002; Eglinger et al. 2017, Fig. 2b). These magmas formed (i) by extraction of juvenile melt
375 from a depleted mantle source that had been metasomatised in an arc setting, and (ii) by
376 reworking of pre-existing Archean crust (Eglinger et al. 2017). This plutonism was broadly
377 coeval with a resurgence of juvenile mafic-ultramafic volcanism (Niandan and Kéniéro
378 volcanic rocks) along the northwest margin of the Kénéma-Man Domain at ca. 2095 Ma

379 (Tegyey and Johan 1989; Milési et al. 1989; Berring et al. 1998; Lahondère et al. 2002;
380 Mériaud et al. 2020). This late episode of volcanism is referred to as the Bandamian Cycle
381 (Hirdes et al. 1996; Mériaud et al. 2020) and has received little attention to date, but likely
382 also includes ca. 2080-Ma-old intermediate to felsic volcanism in the eastern Kédougou-
383 Kénieba Inlier (Allibone et al. 2020). Eruption of juvenile mafic-ultramafic volcanic rocks
384 through crust thickened during the earlier docking of the Archean and Paleoproterozoic
385 domains is inferred to reflect upwelling of the asthenosphere and associated extension caused
386 by litho- spheric delamination or slab breakoff (Eglinger et al. 2017; Mériaud et al. 2020).
387 Further voluminous high-K calc-alkaline and peraluminous granites were emplaced between
388 ca. 2100 and 2080 Ma across much of the southern and western Baoulé- Mossi Domain, but
389 such rocks are absent from northern and eastern parts of the domain. These rocks intrude
390 sedimentary
391 rocks of the Sunyani-Comoé series in southwest Ghana (Appouasso, 2081 ± 1 Ma and
392 Kawtiago, 2088 ± 1 Ma, Hirdes et al. 1992), the area along the Greenville- Ferkessédougou-
393 Bobo-Dioulasso shear zone in Côte d'Ivoire (Férké, 2094 ± 6 Ma, Doumbia et al. 1998),
394 much of southern Mali between the Banifing and Seikorole shear zones (Parra-Avila et al.
395 2017, 2019; Wane et al. 2018), and sedimentary rocks of the Siguiiri (Maléa, 2089 ± 12 Ma,
396 Parra- Avila et al. 2016), Kofi (Balangouma, 2070 ± 6 Ma, Allibone et al. 2020), and Dialé-
397 Daléma series (Saraya, 2079 ± 2 Ma, Hirdes and Davis 2002) in southern Mali and Senegal
398 respec- tively. Together, these high-K calc-alkaline and peraluminous granites form a major
399 magmatic province within the southern and western parts of the Baoulé-Mossi Domain with
400 the most intense plutonism during this period focussed along the boundary between the
401 Archean Kénéma-Man and Paleoproterozoic Baoulé-Mossi domains in Guinea and
402 throughout the Paleoproterozoic rocks of central-southern Mali.

403 Peraluminous two-mica leucogranites formed by either partial melting of metasedimentary
404 source rocks (Egal et al. 2002), and/or older calc-alkaline granites (Tapsoba et al. 2013), with
405 both scenarios involved the reworking of older Baoulé-Mossi Domain crust. The regional
406 extent of the high- K calc-alkaline granites suggests asthenospheric upwelling and
407 lithospheric delamination extended throughout the south- ern and western Baoulé-Mossi
408 Domain, well beyond the boundary with the Archean Kénéma-Man Domain. Extension and
409 unroofing of deep crustal rocks in the Sefwi belt at ca. 2075 Ma (McFarlane et al. 2019a)
410 may also have been another expression of this asthenospheric upwelling and associated
411 lithospheric delamination.

412

413

414 **Space-time-integrated gold metallogeny of the Baoulé-Mossi Domain**

415 Gold deposits discussed in this section are divided into Eoeburnean, Early Eburnean, and
416 Late Eburnean groups, which formed between ca. 2200 and 2135 Ma, ca. 2110 and 2095 Ma,
417 and ca. 2095 and 2060 Ma, respectively (Fig. 3). The characteristics of deposits in each group
418 reflect the tec- tonic, metamorphic, and magmatic processes prevailing across the Baoulé-
419 Mossi Domain during each of these three periods. Goldfarb et al. (2017) provide a
420 comprehensive re- view of the gold endowment of the Baoulé-Mossi Domain whereas
421 Thébaud et al. (2020) review the character, grade, and tonnage for gold deposits discussed
422 hereafter. Gold de- posits referred to as “orogenic” and “intrusion-related” share many
423 features in common with those discussed by Groves et al. (1998) and Goldfarb et al. (2001,
424 2005), Sillitoe and Thompson (1998)m and Lang and Baker (2001), respectively.

425

426 ***Eoeburnean gold deposits (ca. 2200–2135 Ma)***

427 The structural setting, crosscutting relationships with adjacent dated plutonic rocks, and
428 limited radiometric dating of hydro- thermal minerals indicate some orogenic gold deposits in
429 the eastern and central parts of the Baoulé-Mossi Domain formed during the Eoeburnean
430 period. These Eoeburnean deposits (Fig. 3) contain ~10% of the total endowment of the
431 Baoulé- Mossi Domain (Thébaud et al. 2020).

432 At Kiaka, ~4.5 Moz of gold are hosted by amphibolite- facies biotite-kyanite-garnet-
433 amphibole-bearing schists within the Tenkodogo-Fada N’Gourma Domain in southern
434 Burkina Faso (Fontaine et al. 2017; Gauriau et al. 2020). The early disseminated gold event at
435 Kiaka occurred during sinistral- reverse displacement along the NNE-striking Markoye shear
436 zone at 2157 ± 24 Ma (Fontaine et al. 2017), broadly coeval with region-wide lower
437 amphibolite-facies metamorphism and intrusion of a syn-kinematic diorite at 2140 ± 7 Ma are
438 from Markwitz et al. (2016). The geology legend is identical to that used in Fig. 1 (Fontaine
439 et al. 2017). An early orogenic gold event within the ~8 Moz Mana gold district in western
440 Burkina Faso oc- curred contemporaneously with intrusion of the syn-kinematic 2172 ± 6 Ma
441 Wona-Kona and 2176 ± 8 Ma Siou plutons during E-W-directed Eoeburnean shortening and
442 regional up- per greenschist-facies metamorphism (Augustin et al. 2017). The ~0.75 Moz
443 Julie gold deposit in NW Ghana is focussed along E-W-striking structures that developed
444 between ca. 2145 and 2130 Ma (Amponsah et al. 2016; Salvi et al. 2016b), broadly coeval
445 with regional amphibolite-facies metamorphism at 2137 ± 8 Ma (Block et al. 2016). Early
446 orogenic gold mineralisation at the ~10 Moz Wassa deposit in SW Ghana (Perrouy et al.
447 2015) occurred after emplace- ment of the host porphyry at 2191 ± 6 Ma but before emplace-
448 ment of the crosscutting Benso granodiorite at 2157 ± 5 Ma (Parra-Avila et al. 2015). Pyrite
449 grains aligned and stretched within the early fabric at Wassa host gold particles (Perrouy et
450 al. 2016) crystallised at 2164 ± 14 Ma (Le Mignot et al. 2017b). The geometry of the Wassa
451 deposit is controlled by a kilometre-scale NE-striking upright synform (Fig. 4a, b), which

452 affected rocks of both the Sefwi volcanic series and the Tarkwa molasses series. The S1
453 foliation and co-planar gold-bearing quartz veins, which host most of the gold re-
454 source, are both folded by the Wassa synform (Perrouy et al. 2016).

455 Gold skarns also formed locally during the latter part of the Eoeburnean period. The ~3.8
456 Moz Tongon gold deposit (Fig. 3) is confined to a corridor of Eoeburnean orogen-oblique
457 transfer faults that link orogen-parallel N- to NNE-striking reverse faults in the Senoufo belt
458 of northern Côte d'Ivoire. The Tongon deposit formed between 2139 ± 21 Ma and 2128 ± 21
459 Ma (Lawrence et al. 2017), before emplacement of the post-mineralisation Tongon
460 granodiorite at 2139 ± 21 Ma (Lawrence et al. 2017). Despite these imprecise dates, cross-
461 cutting relationships have been used to suggest that the ore- forming process occurred during
462 the culmination of Eoeburnean deformation contemporaneously with peak meta-
463 morphism in northwest Ghana and western Burkina Faso at ca. 2140 Ma (Lawrence et al. 2017).

464 Early Eburnean gold deposits (ca. 2110–2095 Ma)

465 Both orogenic and paleoplacer deposits formed during the Early Eburnean period, and host
466 50 to 70% of the total gold endowment of the Baoulé-Mossi Domain (Thébaud et al. 2020).
467 This particularly well-endowed period of gold mineralisation appears more intense in the
468 eastern part of the Baoulé-Mossi Domain (Fig. 3). Early Eburnean mineralisation is
469 dominated by shear zone and sedimentary rock-hosted orogenic gold deposits typified by the
470 >60 Moz Obuasi deposit (Blenkinsop et al. 1994; Oberthür et al. 1994; Allibone et al. 2002a;
471 Fougereuse et al. 2017; Oliver et al. 2020), and the >40 Moz conglomerate-hosted
472 paleoplacer de- posits of the Tarkwa district (including ~7 Moz at Damang, ~24 Moz at
473 Tarkwa, and ~8 Moz at Iduapriem), both of which are significant parts of the total gold
474 endowment of the Baoulé-Mossi Domain (Sestini 1973; Milési et al. 1991; Eisenlohr and
475 Hirdes 1992; Hirdes and Nunoo 1994; Pigois et al. 2003; Tunks et al. 2004).

476 Gold-bearing quartz pebble conglomerates from the Banket series at Tarkwa (Main and
477 Western Reefs, Fig. 5a) and lower grade reefs at Damang (Fig. 5b) were deposited in a
478 fluvial/lacustrine setting in a larger foreland basin developed on Eoeburnean Ashanti belt
479 basement (Hirdes and Nunoo 1994; White 2011). The ages of detrital zircons in the Banket
480 series and crosscutting intrusions indicate the Banket series was deposited between ca. 2110
481 and 2095 Ma (Davis et al. 1994; Hirdes and Nunoo 1994; Oberthür et al. 1998; Pigois et al.
482 2003; Perrouty et al. 2012), broadly coeval with peak Eburnean deformation and plutonism
483 within the under- lying Ashanti belt and adjacent Kumasi basin (Oberthür et al. 1998).
484 Deposition of gold-bearing conglomerates at Tarkwa was broadly synchronous with
485 Eburnean crustal thickening and exhumation of the adjacent Kumasi series. Thébaud et al.
486 (2020) have suggested that both orogenic gold deposits which formed during the early stages
487 of inversion of the Kumasi series as well as older Eoeburnean gold deposits such as Wassa
488 could have contributed gold to the Tarkwa paleoplacer deposits. This interpretation remains
489 controversial as $\delta^{13}\text{C}_{\text{CO}_2}$ values of high-density CO_2 -rich, water-poor inclu- sions hosted in
490 quartz pebbles from the paleoplacer deposits at Tarkwa differ considerably from the
491 $\delta^{13}\text{C}_{\text{CO}_2}$ values of similar high-density CO_2 -rich inclusions in vein quartz from the giant
492 Obuasi gold system (Klemd et al. 1993; Lüders et al. 2015).

493 At Obuasi and other similar deposits haloes of auriferous arsenopyrite up to 40 m wide
494 developed around mineralised reverse shear zones and along isoclinal fold hinges (Fig. 6a, b).
495 The early Eburnean gold event at Obuasi post-dated the intrusion of the Kumasi series host
496 sedimentary rocks by several small granite plutons at ca. 2105 Ma, with mineralisation
497 occurring at 2098 ± 7 Ma (Oberthür et al. 1998), synchronous with region-wide greenschist-
498 facies meta- morphism at 2092 ± 3 Ma (Oberthür et al. 1998). These age constraints indicate
499 that regional greenschist-facies metamor- phism of the Kumasi series was coeval with gold
500 mineralisation at ca. 2095 Ma (Fougerouse et al. 2017; Oliver et al. 2020). Ongoing

501 shortening of the Kumasi series, or subsequent transpression in the final stages of basin inver-
502 sion, led to the formation of other gold deposits elsewhere along both margins of the Kumasi
503 basin (~17 Moz Prestea- Bogoso gold deposit, Appiah 1991; Allibone et al. 2002b), and shear
504 zones within the Kumasi basin (~0.5 Moz Pampe gold deposit, Salvi et al. 2016a). Deposit-
505 scale structural rela- tionships at the ~7.1 Moz Edikan (Tourigny et al. 2018), the ~8.4 Moz
506 Akyem (Newmont annual report 2017), the ~4.9 Moz Chirano (Allibone et al. 2004), and the
507 ~6.8 Moz Obotan-Esasse (Asanko Mining annual report 2018) gold de- posits (Fig. 3)
508 suggest a similar timing of mineralisation to that at Obuasi.

509 Several orogenic gold deposits in Burkina Faso and NW Ghana also formed at this time. The
510 younger phase of mineralisation in the ~7.8 Moz Mana gold district in western Burkina Faso
511 was largely coeval with sinistral displacement along the NNE-striking Wona-Kona shear zone
512 and intrusion of the Kokoï diorite at 2090 ± 8 Ma (Augustin et al. 2017). Fabric-mineral-vein
513 crosscutting relationships and the age of the host rock sequence at the ~1.9 Moz Inata deposit
514 in the Bouroum belt of northern Burkina Faso indicate that gold deposition occurred after ca.
515 2120 Ma (McCuaig et al. 2016), whereas Re-Os dating of pyrite implies gold mineralisation
516 occurred at 2090 ± 29 Ma (Fougerouse 2011). The ~6.5 Moz Essakane gold deposit in
517 northern Burkina Faso formed during sinistral transpressional reactiva- tion of the Markoye
518 shear zone during the onset of the Eburnean orogeny (Tshibubudze and Hein 2016). The
519 second mineralising event at Kiaka occurred during retrogression from lower amphibolite-
520 facies to upper greenschist-facies conditions coeval with the reactivation of the Markoye
521 shear between ca. 2125 and 2090 Ma, whose timing is bracketed by the emplacement of pre-
522 and post-mineralisation granites (Castaing et al. 2003; Fontaine et al. 2017). The age of the
523 orogenic gold overprint on the Gaoua Eoeburnean porphyry- Cu system is not known but
524 may have occurred between ca. 2100 and 2090 Ma, during the waning stages of the Eburnean
525 tectono-thermal event in Burkina Faso (Le Mignot et al. 2017a). The ~0.1 Moz Bepkong gold

526 deposit in the Wa- Lawra belt of NW Ghana is associated with structures that developed
527 during sinistral displacement along the first-order Jirapa shear zone at ca. 2100 Ma implying
528 it also formed during the Early Eburnean period (Amponsah et al. 2016; Block et al. 2016).
529 Orogenic gold mineralisation, comparable in age and style with that in the eastern part of the
530 Baoulé-Mossi Domain, is much rarer in the western part. Part of the gold resource at the
531 ~13.5 Moz Siguiri deposit in Guinea may have developed during the Early Eburnean period,
532 albeit close to its end, after the switch from compressional to strike-slip tectonics at ca. 2095
533 Ma in the western part of the Baoulé-Mossi Domain (Lebrun et al. 2017). The maximum age
534 for gold mineralisation in the Siguiri area is provided by the youngest detrital zircon age
535 population from the Fatoya sedimentary host rocks, which is 2113 ± 5 Ma (Lebrun et al.
536 2016).
537 Other less common styles of gold mineralisation developed during Early Eburnean time.
538 These include the early formed intrusion-related Au-(Sb, Bi, Te) mineralisation at the ~7.5
539 Moz Morila gold deposit within the amphibolite-facies aureole of a tonalite-granodiorite-
540 leucogranite intrusive complex emplaced between 2098 ± 4 Ma and 2091 ± 4 Ma (McFarlane
541 et al. 2011). The association between early mineralisation at Morila, Te-Bi minerals, and
542 leucogranite veinlets imply mineralisation was directly associated with granite magmatism.

543

544 *Late Eburnean gold deposits (ca. 2095–2060 Ma)*

545 Late Eburnean gold deposits formed between ca. 2095 and 2060 Ma host 20 to 40% of the
546 total gold endowment of the Baoulé-Mossi Domain (Thébaud et al. 2020). Gold
547 mineralisation during the late Eburnean was broadly coeval with emplacement of voluminous
548 high-K granites (Eglinger et al. 2017; Allibone et al. 2020) and eruption of mafic- ultramafic
549 volcanic rocks along the Baoulé-Mossi – Kénéma-Man margin (Mériaud et al. 2020).
550 Although many gold deposits that formed during this period are comparable with orogenic

551 gold deposits, some also have characteristics that imply they formed in aureoles related to
552 high-K granites (Allibone et al. 2020; Lambert-smith et al. 2020). This is par- ticularly
553 apparent in the Kédougou-Kéniéba Inlier where gold mineralisation in the ~17 Moz Loulo
554 district and the ~12 Moz Sadiola district occurred between ca. 2090 and 2060 Ma coe- val
555 with emplacement of nearby high-K granites, but after basin inversion and metamorphism of
556 the host Kofi series (Masurel et al. 2017a; Allibone et al. 2020). In the Loulo district, gold
557 mineralisation was also broadly coeval with iron-(gold) endo- and exo-skarn formation (Fig.
558 7a) along pluton margins within the ca. 2090–2070 Ma Falémé batholith and adjacent Kofi
559 series sedimentary rocks (Schwartz and Melcher 2004; Delor et al. 2010; Lawrence et al.
560 2013a, 2013b; Masurel et al. 2017c; Allibone et al. 2020). Stable isotope (B, C, O) data,
561 however, imply that hydrothermal minerals within these gold deposits were largely derived
562 from metamorphosed calcareous, Na-B-rich evaporitic, and siliciclastic sedimentary rocks
563 within the Kofi series rather than from the skarns or associated granites (Lambert-Smith et al.
564 2016, 2020). Only the gold prospects within the imme- diate vicinity of the dyke swarm
565 along the eastern margin of the Falémé batholith have isotopic characteristics indicative of
566 the direct input of magmatic fluid during mineralisation (Lambert-Smith et al. 2020).

567 In southern Côte d’Ivoire, the ~2.5 Moz Bonikro gold de- posit (Ouatarra 2015) includes
568 early intrusion-related gold mineralisation (Fig. 7b) formed at 2086 ± 4 Ma (Masurel et al.
569 2019) and later orogenic gold mineralisation formed at 2074 ± 16 Ma (Re-Os on
570 molybdenite, Masurel et al. 2019). At Morila, the Early Eburnean intrusion-related gold (Bi,
571 Sb, Te) mineralisation was overprinted by orogenic gold (As) mineralisation at 2074 ± 14 Ma
572 (McFarlane et al. 2011). At both Morila and Bonikro, the later orogenic gold event ac- counts
573 for the bulk of the gold resource.

574 Orogenic gold mineralisation that occurred during the later part of the Eburnean orogeny (ca.
575 2095–2060 Ma) formed along retrograde P-T-t paths, and was associated with either local

576 transpression, wrench, or transtensional deformation (Goldfarb et al. 2017). These relatively
577 minor, small- displacement and/or low-strain deformation events appear re- stricted to small
578 parts of the Baoulé-Mossi Domain, and likely cannot be correlated across the whole region.
579 Orogenic gold deposits that formed during this period may include the later mineralising
580 event at the Obuasi deposit (Fougerouse et al. 2017; Oliver et al. 2020), gold mineralisation at
581 the major Ahafo deposit in western-most Ghana (Baah-Danso 2011; Masurel et al. 2021), the
582 Siguiri deposit in Guinea (Lebrun et al. 2017), and both mineralising events at the Kalana de-
583 posit in southern Mali (Salvi et al. 2016c).

584 The second mineralising event at the Obuasi deposit in- cludes some particularly high-grade
585 parts of the resource (Fig. 8a) that formed during late transpressional reactivation of segments
586 of the Obuasi and Main Reef shear zones, both older contractional structures active during
587 earlier basin inversion. New mineralised faults also formed at this time, including the Ashanti
588 Fissure at the northern end of the deposit and several minor faults in various prospects south
589 of the deposit (Allibone et al. 2002a; Fougerouse et al. 2017; Oliver et al. 2020). It remains
590 unclear whether this gold was remobilised from rocks mineralised during the Early Eburnean
591 phase, or introduced during late Eburnean transpressional deformation (Fougerouse et al.
592 2016). Some mineralised lodes in other deposits along the south- east margin of the Kumasi
593 basin, such as the ~17 Moz Bogoso-Prestea gold system (Fig. 8b), are developed in similar
594 structural settings to the later mineralised rocks at Obuasi (Allibone et al. 2002b). Gold
595 mineralisation in the ~10 Moz Ahafo district formed during transpressional sinistral-reverse
596 reactivation of the Kenyase-Yamfo fault system (Baah-Danso 2011; McFarlane et al. 2019a)
597 after exhumation of high-grade metamorphic rocks in the north- west part of the Sefwi belt at
598 ca. 2073 Ma (McFarlane et al. 2019a) and emplacement of ca. 2080 Ma two-mica granites in
599 the adjacent Sunyani-Comoé basin (Hirdes et al. 2007; Jessell et al. 2012). Gold
600 mineralisation in the ~13.5 Moz Siguiri district in Guinea occurred during transpression and

601 transient transtension after region-wide compressional de- formation at ~2095 Ma (Lebrun et
602 al. 2017). At the ~2 Moz Kalana deposit further east of the Siguiri basin, gently dip- ping
603 auriferous quartz veins localised along thrust faults and near vertical auriferous quartz veins
604 in extensional fractures crosscut the regional penetrative schistosity and were interpreted to
605 have both formed during progressive E-W- directed shortening (Salvi et al. 2016c). Gently
606 dipping gold-bearing quartz veins similar to those at Kalana are rel- atively rare in the
607 Baoulé-Mossi Domain, but have also been described at the Obotan deposit in the Kumasi
608 basin and the Damang deposit in the Ashanti belt (Tunks et al. 2004). The orogenic gold
609 event that overprinted paleoplacer gold at Damang is hosted within a contractional sub-
610 horizontal fault-fracture mesh and associated quartz vein arrays (Tunks et al. 2004; White
611 2011; White et al. 2015, Fig. 9a, b), whose associated alteration haloes include hydro-
612 thermal xenotime dated at 2063 ± 9 Ma (Pigois et al. 2003). Late gold mineralisation at
613 Wassa, dated at 2055 ± 18 Ma (Re-Os on pyrite, Le Mignot et al. 2017b), is hosted by quartz
614 veins that crosscut the earlier foliation which hosts auriferous pyrite (Perrouy et al. 2015).

615 At Obuasi, white mica K-Ar cooling ages of ca. 2080 Ma (Oberthür et al. 1998) suggest gold-
616 related hydrothermal ac- tivity had ceased by this time, whereas the younger ages ob- tained
617 from hydrothermal minerals at the Damang and Wassa deposits indicate hydrothermal
618 activity continued for a further 10 to 20 myr in these locations. Such differences in timing
619 reflect the progressive waning of orogenic activity after ca. 2095 Ma in the eastern part of the
620 Baoulé-Mossi Domain and after ca. 2085 Ma in its western part (Parra-Avila et al. 2017;
621 Grenhölml et al. 2019). These data indicate most, if not all, gold deposits in the Baoulé-Mossi
622 Domain had formed by ca. 2060 Ma, with locally reported deformation after ca. 2060 Ma
623 reflecting the far-field effects of orogenic activity beyond the current limits of the Baoulé-
624 Mossi Domain such

625 as Orisian (ca. 2000 Ma) deformation and plutonism in parts of northeast South America.

626

627

628

629 **Discussion**

630

631 **Geodynamic context of gold mineralisation**

632

633 The north-central part of the Baoulé-Mossi Domain hosts sev-
634 eral minor porphyry-Cu and VMS base metal occurrences as well as the much more widespread gold deposits described
635 above (Fig. 3). The distribution and age of these porphyry-
636 Cu and VMS occurrences provide additional clues as to the tectonic context of the formation of nearby gold deposits of broadly
637 the same age. In Burkina Faso, porphyry-Cu mineralisation at Gaoua was broadly coeval with
638 VMS mineralisation at Perkoa, Wayen, Tiébélé, and Poura (Figs. 10a and 11). The age of the
639 Perkoa VMS (Zn-Ag) deposit is constrained between that of the 2171 ± 7 Ma host rhyolite
640 (Le Métour et al. 2003) and a crosscutting quartz-diorite emplaced at 2175 ± 1 Ma (Schwartz
641 and Melcher 2003). VMS (Zn-Pb-
642 Cu-Ag) mineralisation at Tiébélé (Ilboudo et al. 2017) is hosted by 2156 ± 9 Ma rhyolite (Kaboré 2004), whereas gold-rich VMS mineralisation at
643 Poura remains undated but is inferred to have formed at a similar time to the gold-poor VMS
644 occurrences in the region. Porphyry-Cu mineralisation at Gaoua occurred at 2165 ± 24 Ma
645 (Le Mignot et al. 2017a), an age whose uncertainty brackets the emplacement ages of the
646 2185 ± 12 Ma host Diénémera hornblende-diorite and 2172 ± 9 Ma post-mineralisation
647 Gongondy gabbro (Brownscombe 2009). The ages and spatial distribution of the Gaoua
648 porphyry-Cu and Perkoa-style VMS occurrences suggest western Burkina Faso was located
649 in an arc/back-arc setting (Seedorff et al. 2005; Franklin et al. 2005) throughout much of
650 Eoeburnean time between ca. 2180 and 2140 Ma (Figs. 10a and 11).

651 Eoeburnean orogenic gold mineralisations such as Kiaka (I), Julie, and Wassa (I) are
652 interpreted to have formed during the inversion of this inferred back-arc rift (Figs. 10a and
653 11). The paucity of Eoeburnean gold deposits formed over this period may reflect the
654 following: (1) the restricted extent of rocks of this age, which are largely confined to the
655 eastern part of the Baoulé-Mossi Domain; (2) the higher metamorphic grade and lack of
656 preservation of prospective upper to mid- crustal rocks in parts of northwest Ghana and
657 eastern Burkina Faso; and (3) reworking of dehydrated, un-prospective felsic Archean
658 material by calc-alkaline plutons in south-eastern Ghana during accretion of juvenile
659 Paleoproterozoic volcano-sedimentary rocks onto blind deep crustal Archean basement east
660 of the Baoulé-Mossi Domain at ca. 2140 Ma (De Kock et al. 2011; Petersson et al. 2016).
661 This Archean basement, which is distinct from the Kénéma-Man nucleus, is inferred to occur
662 in southeast Ghana and Togo based on xenocrystic zircon cores (De Kock et al. 2011), Lu-Hf
663 data (Petersson et al. 2016, 2018), and the presence of Archean detrital zircons in Eburnean
664 sedimentary rocks (Pigois et al. 2003; Grenhölml et al. 2019).

665 Giant paleoplacer gold deposits are generally considered to form in retro-arc foreland basins,
666 inboard of the accretionary system where detrital gold is transported, deposited, and reworked
667 (Solomon and Groves 1994). Gold-bearing con- glomerates were deposited within the
668 Tarkwa basin as it was deformed within an actively accreting margin (Figs. 10b and
669 11) (Milési et al. 1991; Hirdes and Nunoo 1994; Pigois et al. 2003; White et al. 2011, Le
670 Mignot et al. 2017b). Eoeburnean gold deposits such as those in the Ashanti belt in southwest
671 Ghana and the Ouadalan-Gouroul belt and Fada N’Gourma Domain in Burkina Faso have
672 been suggested as the sources of gold in the Tarkwa paleoplacer deposit (Klemd et al. 1993;
673 Perrouty et al. 2015) and debated in Lüders et al. (2015). However, given the protracted
674 nature of metamorphism, gold mineralisation, and associated deformation along the eastern
675 margin of the Kumasi basin (Eisenlohr and Hirdes 1992; Fougrouse et al. 2017; Oliver et al.

676 2020), Early Eburnean orogenic gold deposits within metasedimentary rocks of the Kumasi
677 series (Obuasi, Edikan, Obotan-Easse) may also be viable alternative sources for the
678 Tarkwaian paleoplacer gold deposits (Thébaud et al. 2020). The greater volume of shallow
679 water to subaerial series in the eastern part of the Baoulé- Mossi Domain (Tarkwa, Sestini
680 1973; Pigois et al. 2003; Sambragian, Lüdtkke et al. 1999; Safane-Téhini, Bossière et al. 1996;
681 Baratoux et al. 2011) implies different amounts of exhumation and denudation across the
682 Baoulé-Mossi Domain at ca. 2110 Ma (Ledru et al. 1994; Feybesse et al. 2006).

683 The relative proportions of orogenic gold deposits formed during the early and late Eburnean
684 phases is uncertain because of the following: (1) the age of some major deposits remains
685 unknown (Syama); (2) the relative proportions of gold precipitated during different events
686 in multistage deposits is poorly defined (Obuasi); and (3) the amount of gold eroded and
687 reworked during the Eburnean orogenic activity is unknown but likely exceeded 10s of Moz
688 given the amount of gold in Tarkwa paleoplacer deposits. Major late Eburnean orogenic gold
689 deposits have been discovered across the southern and western Baoulé-Mossi Domain
690 whereas early Eburnean orogenic gold deposits appear restricted to the eastern part of the
691 Baoulé-Mossi Domain (Milési et al. 1992; Le Mignot et al. 2017b; Goldfarb et al. 2017 and
692 references therein). Many deposits either formed (Loulo, Sadiola) or were upgraded (Obuasi)
693 during late Eburnean tectono-thermal activity between ca. 2095 and 2060 Ma, soon after the
694 change from orogen-perpendicular shortening to prolonged, diachronous, multistage
695 transpressional to transtentional tectonics at ca. 2090 Ma (Figs. 10c and 11). A 10–20°
696 clockwise rotation in the bulk shortening direction from NW-SE to NNW-SSE occurred
697 during the late Eburnean. This rotation coupled with the generally steep dip of layering across
698 much of the Baoulé- Mossi Domain resulted in major structures being unfavourably oriented
699 for dip-slip displacement. Stress field switches such as this have been proposed as a

700 mechanism that promoted permeability and fluid flow, thus driving the forma-
701 tion of larger gold deposits (Ledru et al. 1991; Milési et al. 1992; Lebrun et al. 2017).
702 Late Eburnean delamination of a lithospheric root and/or the roll back of a subducted slab
703 along the suture between the Archean Kénéma-Man Domain and Paleoproterozoic Baoulé-
704 Mossi Domain has been cited as a cause of lithospheric instabil- ity and high heat/fluid flux
705 in the crust after ca. 2095 Ma, leading to widespread orogenic and local intrusion-related gold
706 mineralisation in the southern and western parts of the Baoulé- Mossi Domain (Eglinger et al.
707 2017). Alternatively, late Eburnean granitic plutonism and orogenic gold mineralisation could
708 also reflect the strike-slip reactivation of trans- lithospheric faults and lithostratigraphic
709 boundaries during lateral escape of the crust, which induced decompression melting and
710 fluid migration from deeper to shallower levels in the crust (Masurel et al. 2017c; Grenholm
711 et al. 2019). The formation of intrusion-related gold deposits in the Baoulé-Mossi Domain
712 ap- pears to span much of Eburnean orogeny from ca. 2098 Ma (Morila) through to ca. 2074
713 Ma (Bonikro). Intrusion-related gold deposits worldwide typically formed in back-arc
714 settings or on deformed accretionary margins (Sillitoe and Thompson 1998; Lang and Baker
715 2001, Fig. 10c).

716

717 *Insights from the Guiana Shield in South America*

718 Although the Paleoproterozoic rocks of the Guiana Shield have not yet received the same
719 research as the Baoulé-Mossi Domain, Vanderhaeghe et al. (1998), Milési et al. (2003), Delor
720 et al. (2001, 2003), Kioe-A-Sen et al. (2016), Kroonenberg et al. (2016), and Tedeschi et al.
721 (2020) offer insights into the tectono-metallogenic evolution of these Paleoproterozoic rocks,
722 which can be compared with that of the Baoulé-Mossi Domain. In French Guiana, an early
723 gold event was associated with a period of crustal growth and mag- matic accretion,
724 following the production of an older basaltic crust. A second orogenic gold event coeval with

725 deposition of gold paleoplacers occurred during a period of collision and crustal reworking.
726 This history is similar to that of the Baoulé-Mossi Domain.
727 Although two or more pulses of Paleoproterozoic gold mineralisation appear to have occurred
728 within the Guiana Shield, limited data suggests many of the discovered gold deposits formed
729 during sinistral shearing along major crustal discontinuities, opening of pull-apart basins, and
730 leucogranite emplacement late in the Transamazonian orogeny between ca. 2110 and 2060
731 Ma (Vanderhaeghe et al. 1998; Delor et al. 2001, 2003; Milési et al. 2003). Precise absolute
732 ages on the timing of gold and hydrothermal events within the Guiana Shield are scarce. At
733 the Dorlin VMS deposit, overprinting orogenic gold mineralisation occurred at 2067 ± 1 Ma
734 (Marcoux and Milési 1993). Hydrothermal titanite at the Karouni orogenic gold deposit in
735 Guyana crystallised at 2084 ± 14 Ma (Tedeschi et al. 2018). These dates correlate closely with
736 the major late Eburnean period of orogenic gold mineralisation in West Africa. Younger gold
737 mineralisation occurred after ca. 2030 Ma in the Guiana Shield (Tedeschi et al. 2018),
738 whereas no gold mineralisation as young as this has been reported from the Baoulé-Mossi to
739 date. The geological context of this younger mineralisation within the Guiana Shield
740 remains unclear.

741

742 *Comparison with the Neoproterozoic Yilgarn and Superior Cratons*

743 Some aspects of the geological history, metallogenic evolution, and gold endowment of the
744 Baoulé-Mossi Domain closely resemble those of the Neoproterozoic Superior (Canada) and
745 Yilgarn (Australia) cratons (Robert et al. 2005). Orogenic cycle(s) in each region included a
746 change from early mafic volcanism to later siliciclastic and volcanoclastic sedimentation
747 above unconformities, followed by bulk incremental deformation, peak regional
748 metamorphism, and granitic plutonism. Key geological features of the three most richly

749 endowed gold subprovinces in these cratons: Ashanti, Abitibi, and Kalgoorlie, and compared
750 in Fig. 12.

751 Waning volcanic activity in these subprovinces corresponded with an increase in clastic
752 sedimentation. The Abitibi subprovince includes sedimentary rocks comparable with the
753 sedimentary series present in the Baoulé-Mossi Domain. The Porcupine Group consists of
754 turbiditic rocks several kilometres thick, which were deposited in a deep water setting. The
755 younger fault-bounded Timiskaming Group consists, in part, of abundant shallow water to
756 subaerial conglomerates (Card and Poulsen 1998). In contrast, the extensive sedimentary
757 series amongst the Paleoproterozoic terranes of the Baoulé-Mossi Domain (Sunyani-Comoé,
758 Kumasi, Siguiri, Kofi, Dialé-Daléma, Grenholm et al. 2019) and Superior Craton (Pontiac,
759 Quetico, English River, Frieman et al. 2017) have no equivalents in the
760 2008; Squires et al. 2010; Czarnota et al. 2010; Vielreicher et al. 2016) cratons with the
761 Paleoproterozoic terranes of the Baoulé-Mossi Domain (this study and references therein)
762 Neoproterozoic Yilgarn Craton. Sedimentary rocks comparable with the deep water series in the
763 Ashanti and Abitibi subprovinces may be represented in the Kalgoorlie subprovince by the
764 Merougil series (Squires et al. 2010). Late orogenic coarse- clastic rock successions in the
765 Kalgoorlie subprovince, such as Kurrawang series (Krapez et al. 2000), are more comparable
766 in sedimentological character and scale with molasse, than deep water rock sedimentary
767 sequences. Plutonism accompanied in- termediate to felsic volcanism and outlasted clastic
768 sedimentation in all three subprovinces. In addition, there is a common evolu- tion from early
769 mafic to intermediate plutons and/or TTGs, to high-K granites and/or two-mica-granites
770 throughout the orogen- ic cycle (in the Yilgarn Craton, Zibra et al. 2020; in the West African
771 Craton, McFarlane et al. 2019a).

772 Despite the difficulty in correlating deformation events from one region to another within the
773 Baoulé-Mossi Domain such as local versus regional deformation and progressive versus

774 multiphase deformation, the three subprovinces experienced a similar history of
775 deformation to that outlined by Robert et al. (2005). Diachronous deformation has been dem-
776 onstrated in all cratons, although this is not necessarily apparent at the belt scale (Robert et al.
777 2005; Czarnota et al. 2010; Parra-Avila et al. 2017). The diversity of gold mineralisation
778 styles, ore and alteration assemblages, and overprinting relationships indicate several gold
779 events often within individual camps, which spanned at least the late stages of crustal growth
780 in all three subprovinces (Robert et al. 2005; Goldfarb et al. 2017; Thébaud et al. 2018). The
781 diachronous nature of gold mineralisation in these cratons reflects the timing of terrane
782 accretion in different supracrustal belts (Robert et al. 2005; Vielreicher et al. 2016).

783 Despite these many similarities, there are also a number of important differences between
784 these Paleoproterozoic and Neoproterozoic subprovinces, which reflect contrasting aspects of
785 their geodynamic evolution and which have important implications for exploration targeting
786 (Fig. 12). Firstly, diachronous gold mineralisation in the Abitibi and Kalgoorlie subprovinces
787 was spread over ~20 to 40 myr, whereas gold mineralisation in the Baoulé-Mossi Domain
788 spanned over ~100 myr. Secondly, komatiitic lava flows and associated nickel ores are absent
789 from the Baoulé-Mossi Domain, a difference that highlights the secular cooling of the Earth
790 asthenosphere between the Neoproterozoic and Paleoproterozoic. Paleoproterozoic Ni-Cu-(PGE)
791 mineralisation is, however, present in western Côte d'Ivoire where it is associated with the
792 Samapleu-Yacouba layered mafic-ultramafic complex, which intruded Archean crust. Mafic-
793 ultramafic layers within the complex consist of lower MgO websterite, pyroxenite, peridotite,
794 and gabbro-norite and were genetically related to the Ni-Cu-(PGE) mineralisation that
795 occurred between 2091 ± 18 and 2080 ± 13 Ma (Gouedji et al. 2014). Thirdly, the tectono-
796 stratigraphic setting of the most richly endowed parts of the three cratons differs. Major gold
797 districts in the Superior and Yilgarn cratons such as Sunrise Dam in the Kalgoorlie
798 subprovince and Val-d'Or in the Abitibi subprovince are hosted in greenstones. In contrast,

799 >90% of known gold endowment in the Baoulé-Mossi Domain occurs either within early
800 Eburnean large-scale sedimentary basins or along fault systems that coincide with their
801 margins (Goldfarb et al. 2017; Oliver et al. 2020). Lastly, the resurgence of mafic to felsic
802 volcanism near the suture between Paleoproterozoic and Archean domains in West Africa
803 occurred at similar times as molasse deposition in foreland basins across the Baoulé-Mossi
804 Domain (Mériaud et al. 2020). Such late volcanism is absent in both the Neoproterozoic Yilgarn
805 and Superior cratons, and this may reflect the intrinsic buoyancy of Archean lithospheric
806 roots due to their high Mg and low Fe contents, which inhibited lithospheric delamination
807 thus preventing the generation of widespread post-orogenic melts (Begg et al. 2009).

808

809 **Conclusions**

810 Gold mineralisation in the Paleoproterozoic Baoulé-Mossi Domain formed during three
811 pulses within a >100-Myr-long period of crustal growth and differentiation. The spatial and
812 temporal distribution of known gold deposits is consistent with accretion-collision tectonics,
813 characterised by a complex interplay between deformation, crustal thickening, burial,
814 metamorphism, partial melting, and exhumation. The most prolific period for ore deposition
815 correlated with the Eburnean orogeny. Changes in the geodynamic setting, which affected the
816 kinematics of Eburnean deformation and the character of associated plutonism, coincided
817 with increases in the flux of gold mineralisation across the Baoulé-Mossi Domain between
818 ca. 2110 and 2060 Ma.

819 This review raises at least one fundamental question. Was the total gold budget acquired right
820 from the earliest geodynamic stages through accretion of juvenile crust and then remobilised
821 during crustal reworking? Or was an initial gold budget refined by mantle-derived magmas
822 throughout the orogenic cycle? The missing link of the West African metallogenic puzzle lies
823 in the apparently Au-poor Archean Kénéma-Man Domain. Future research should determine

824 if the metallogeny of the Paleoproterozoic West Africa was notably different from its Archean
825 counterpart. Furthermore, regions within the Archean Kénéma-Man Domain, about which
826 little is known and which remain largely under-explored, offer prospective targets for eco-
827 nomically significant Archean and/or Paleoproterozoic gold ores (New Liberty, Ndablama,
828 and Kana Hills deposits in Liberia; Baomahun and Kalmaro deposits in Sierra Leone, Foster
829 and Piper 1993; Markwitz et al. 2016).

830

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832

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843

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