

Constraints on the Precambrian paleogeography of West African Craton

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Abstract

This chapter reviews the Precambrian paleogeography of West African Craton (WAC), which is critical for understanding the plate tectonics on early Earth and the configurations of Proterozoic supercontinents. We have summarized the geological history and paleoclimate indicators of WAC, and thoroughly evaluated the available paleomagnetic data. Although we found that WAC has essentially no highly reliable paleomagnetic data of pre-Ediacaran age, ubiquitous mafic dyke swarms could be promising targets for paleomagnetism. In this context, we compiled all high-precision ages pertaining to the mafic intrusions recognized in WAC, and constructed a magmatic barcode for global comparisons. Finally, various paleogeographic reconstruction models regarding the position of WAC in supercontinents Nuna/Columbia and Rodinia were presented and discussed. Because of the lack of robust data, WAC is poorly constrained paleogeographically in Proterozoic supercontinents. We hope this review can provide a basic introduction with regard to the paleogeography of WAC in the Precambrian and advocate future studies.

Keywords: West African Craton; paleomagnetism; supercontinents; LIPs; paleogeography; Precambrian

13.1 Introduction

Paleogeography of cratonic pieces, and their amalgamations, fragmentations, and interactions, are considered to have profoundly influenced the Earth's geological, biological, and environmental cycles. One end-member scenario of paleogeographic evolution is that nearly all cratons assembled together to form supercontinents. Our knowledge regarding the configurations of supercontinents has been greatly improved, owing to increasing geological evidence discovered worldwide, advances in the visualization of kinematic reconstructions, and the inventions and improvements of field and laboratory techniques. For Precambrian time, however, due to less preserved geological records, the configurations of supercontinents and drift histories of individual cratons are still highly debated. As one of the major cratons in the area of exposure (Fig. 13.1A), West African Craton (WAC) is an important component for global paleogeographic reconstructions. Yet, the Precambrian paleogeography of WAC and its relationships with other cratons are very loosely constrained due to the scarcity of robust paleogeographic indicators. In this contribution, we attempt to summarize available reconstruction models pertinent to WAC, including an overview of its tectonic history for intracratonic and global comparisons, a thorough evaluation of all available paleomagnetic data, and an update of its magmatic record. We hope this review could set the stage for future studies regarding the paleogeography of WAC in Precambrian time.

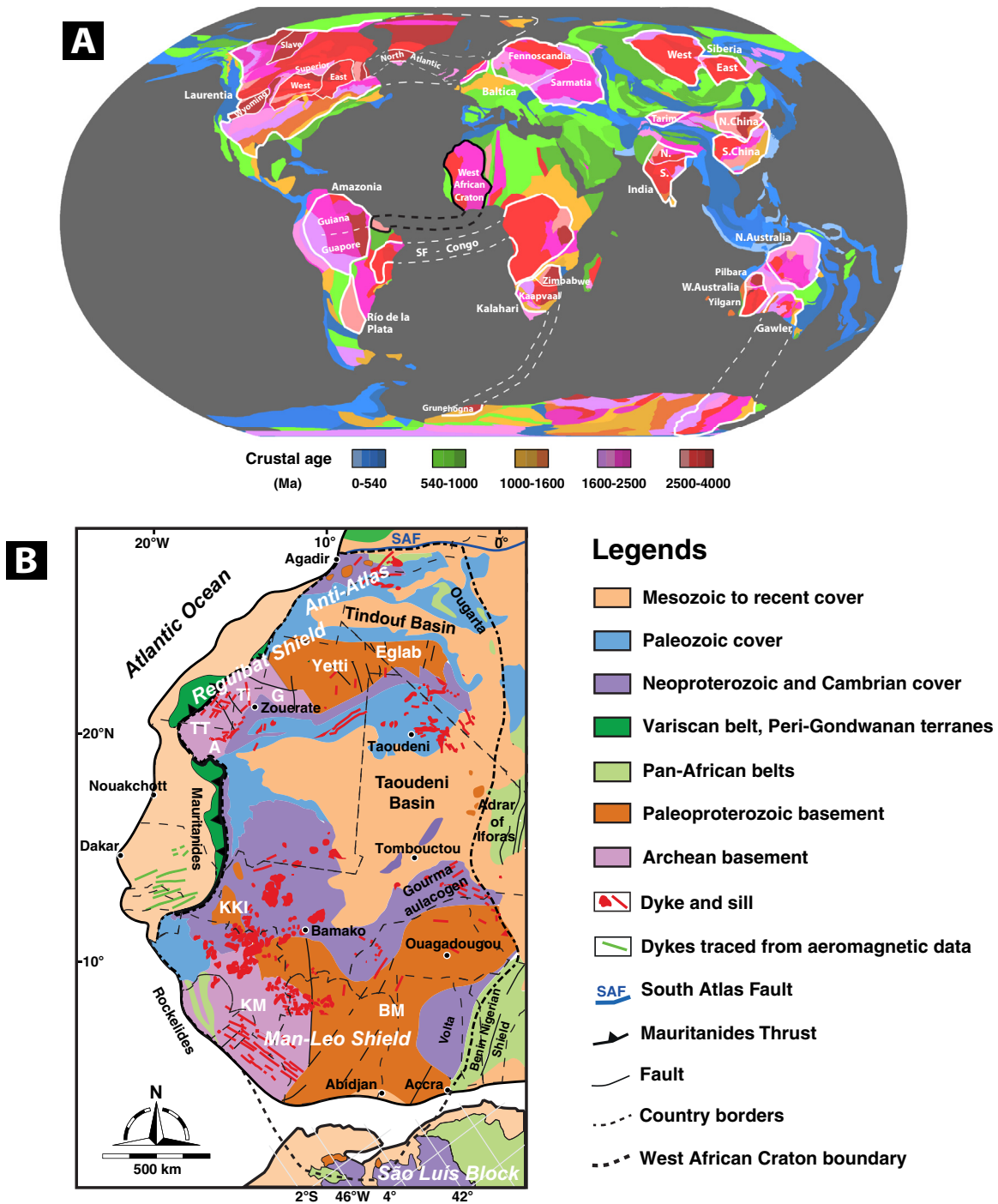


FIGURE 13.1 (A) Global age distribution of continental crust. White lines highlight major cratonic blocks. Black lines delineate the boundary of West African Craton, including the São Luís Block. SL, São Luís; SF, São Francisco. (B) Geological map of West African Craton. TT, Tasiast-Tijirit; A, Amsaga; Ti, Tiris; G, Ghallaman; KKI, Kedougou-Kéniéba and Kayes inliers; KM, Kénéma-Man; BM, Baoulé-Mossi (Ennih and Liégeois, 2008). Modified from Ennih, N., Liégeois, J.P., 2008. The boundaries of the West African craton, with special reference to the basement of the Moroccan metacratonic Anti-Atlas belt. Geological Society, London, Special Publications 297(1), pp. 1–17.

13.2 Geology of West African Craton

WAC is bounded on all sides by Pan-African orogens (Fig. 13.1B) that developed during late Neoproterozoic–Cambrian time. Boundaries are as follows: the South Atlas Fault to the north of the craton, Trans-Saharan Belt to the east, Bassarides/Mauritanides to the west, and to the south a Gondwana-inherited connection to the São Luís Block and Gurupi Belt that are now located in northern Brazil (Ennih and Liégeois, 2008). Basement rocks in WAC are exposed in three regions: Man-Leo Shield in the south, Reguibat Shield in the north, and Anti-Atlas Belt at the northernmost cratonic limit (Fig. 13.1B). Both Man-Leo and Reguibat Shields are characterized by Archean basement in the west and Paleoproterozoic basement in the east. In the Anti-Atlas Belt of Morocco, Paleoproterozoic basement is exposed within a number of inliers, and there is no Archean basement observed. Man-Leo and Reguibat Shields are straddled by the Taoudeni Basin, which was initiated in late Mesoproterozoic time, and was possibly associated with extension caused by continental breakup along the craton's margins (Bertrand-Sarfati et al., 1991; Rooney et al., 2010). In the north, the Tindouf Basin lies between Reguibat Shield and Anti-Atlas Belt; it consists mostly of Phanerozoic sedimentary sequences but the oldest part in the north flank of the basin could potentially be correlated with early Neoproterozoic strata in the Taoudeni Basin (Bertrand-Sarfati et al., 1991). Here, we briefly review the geology of Man-Leo and Reguibat Shields and Anti-Atlas Belt in an attempt to draw temporal and spatial comparisons across WAC.

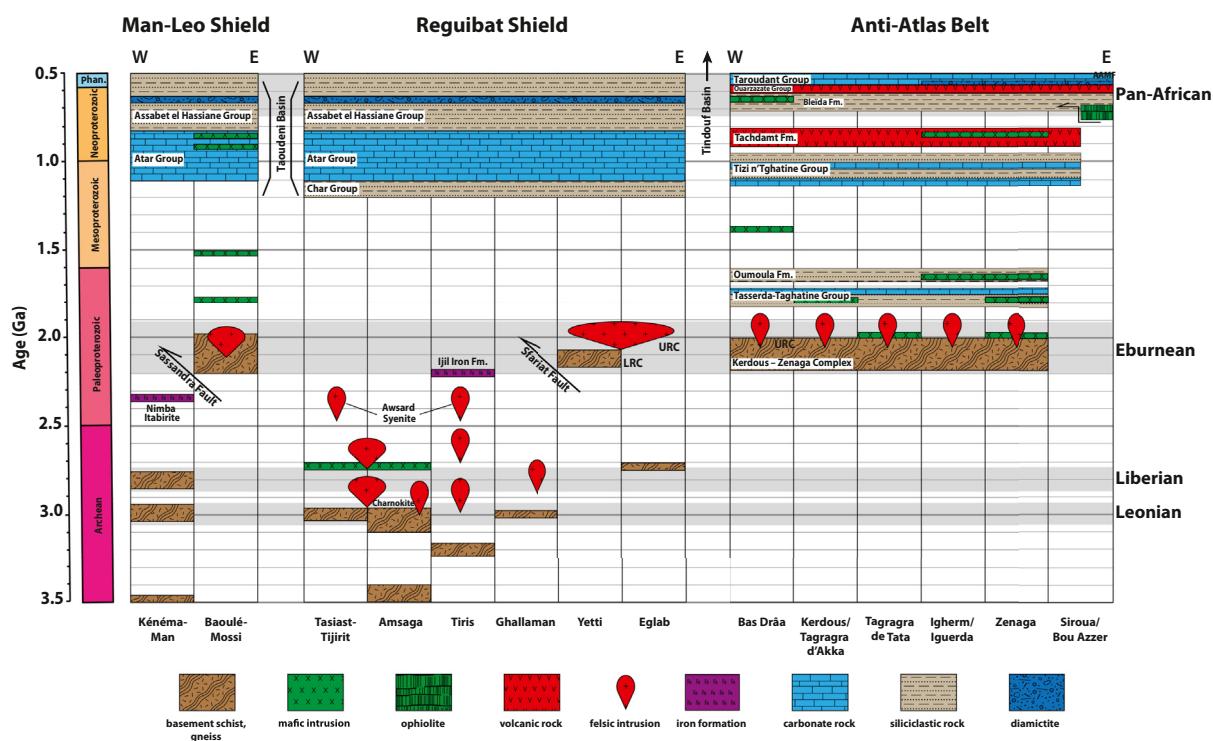


FIGURE 13.2 Time-space diagram demonstrating the spatial and tectonic relationships between the tectonostratigraphic units of West African Craton. LRC, Lower Reguibat Complex; URC, Upper Reguibat Complex; AAMF, Anti-Atlas Major Fault. Neoproterozoic dykes intruded basement rocks in Man-Leo Shield.

13.2.1 Man-Leo Shield

Archean basement of Man-Leo Shield, known as Kénéma-Man domain, crops out in Guinea, Sierra Leone, Liberia, and Ivory Coast (Fig. 13.1B). The oldest exposed rock is the 3.53 Ga granite-gneiss formation identified in Guinea (Thiéblemont et al., 2001). Afterward, continental growth advanced through successive magmatism and accretion at 3.05–2.95 Ga (Leonian orogeny) and 2.85–2.75 Ga (Liberian orogeny) (Fig. 13.2). Paleoproterozoic basement (a.k.a. Birimian terranes) of Man-Leo Shield consists of Baoulé-Mossi domain mainly in Burkina Faso, Ghana, Togo, Ivory Coast, and Mali, the Kedougou-Kéniéba and Kayes inliers in Senegal, and the São Luís Block in Brazil. Paleoproterozoic basement is characterized by volcanic and siliciclastic successions of Birimian

Supergroup, with numerous episodes of intrusions (Grenholm et al., 2019). Collision between Kénéma-Man and Baoulé-Mossi domains occurred at ~ 2.15 Ga through the Eburnean orogeny (Baratoux et al., 2011) along the Sassandra mylonite belt (Fig. 13.2), and was followed by syn- and post-orogenic granitic intrusions (Rocci et al., 1991). Subsequent cratonic stability was punctuated by localized mafic intrusions at 1.79–1.76 Ga, 1.57–1.52 Ga, and 0.91–0.85 Ga (Baratoux et al., 2019).

Epicratonic sedimentation began in the latest Mesoproterozoic in the Taoudeni Basin (Rooney et al., 2010), starting with the basal conglomerate to mark the erosional surface (Fig. 13.2). Subsequently, Char Group is dominated mostly by siliciclastic rocks with minor carbonates and evaporites. Above Char Group is Atar Group, which is characterized by carbonates and shales. Overlying Atar Group is Assabet el Hassiane Group, which mainly consists of siliciclastic successions. Above a regional angular unconformity, a widespread layer of diamictite is (usually) correlated with the c.635-Ma Marinoan glaciation (Fig. 13.2). Atop the diamictite are the mainly siliciclastic Téniaouri and Falaise d'Atar Groups (Bertrand-Sarfati et al., 1991).

The ages of Taoudeni sequences are poorly constrained. Rb-Sr ages on clay minerals suggested that most if not all of the Taoudeni successions were Neoproterozoic (Clauer, 1976; Clauer et al., 1982; Clauer and Deynoux, 1987). But a recent Re-Os geochronological study from the organic-rich sediments in lower Atar Group yielded ages of 1109 ± 22 Ma and 1107 ± 12 Ma (Rooney et al., 2010), which would suggest Char and Atar Groups are ~ 200 Ma older than previously thought. The Mesoproterozoic ages are consistent with relatively muted $\delta^{13}\text{C}$ isotopic compositions (Kah et al., 2012). Mafic dykes, which could possibly be associated with the early-Neoproterozoic continental breakup, intruded Atar Group at 915 ± 7 Ma and 867 ± 16 Ma (Baratoux et al., 2019). The age of Assabet el Hassiane Group is suggested to be mid-Neoproterozoic because its lateral stratigraphic equivalent unit, the Aïoun Group, contains a prominent Tonian (c.880 Ma) detrital-zircon population (Bradley et al., 2015). Above the c.635 Ma Marinoan-equivalent diamictite horizon, tuff beds in Téniaouri Group have been dated by U-Pb SHRIMP on zircon at c.610–605 Ma (see Álvaro et al., 2007).

13.2.2 Reguibat Shield

Archean basement of Reguibat Shield is mainly exposed in Western Sahara, northern Mauritania, and Mali, and is divided into four domains that are separated by faults: Amsaga in the southeast, Tasiast-Tijirit in the southwest, Tiris in the center, and Ghallaman in the northeast (Rocci et al., 1991; Key et al., 2008; Schofield et al., 2012) (Fig. 13.1B). Amsaga contains the 3.5–3.4-Ga crust (Auvray et al., 1992; Potrel et al., 1996) and is thought to represent the oldest crustal growth in the shield (Fig. 13.2). Later on, there was a charnockite intrusion at ~ 3.0 Ga, which marks the major granulite-facies metamorphic event between 3.2 Ga and 3.0 Ga (Potrel et al., 1998; Key et al., 2008). At ~ 2.7 Ga, granitic and gabbroic intrusions were emplaced coevally, and dykes of this age are observed with a NE strike (Auvray et al., 1992; Potrel et al., 1998; Tait et al., 2013). Contemporaneous bimodal intrusions are also found in southern Tasiast-Tijirit across Tâçarât-Inemmaûdene Shear Zone (TISZ). In Tasiast-Tijirit, the earliest crustal materials, manifested by gneisses and granitic intrusions, are dated at ~ 3.0 Ga (Chardon, 1997; Key et al., 2008). Similarly aged granite also intruded westernmost Amsaga near the TISZ (Key et al., 2008). There is a 2.46-Ga syenite in Awsard, northernmost Tasiast-Tijirit (Montero et al., 2014). In Tiris, the major intrusive events occurred at 3.0–2.9 Ga and 2.7–2.6 Ga (Schofield et al., 2012) and the younger 2.46 Ga syenitic suite is also observed (Montero et al., 2014). Massive Ijil Iron Formation is suggested to be ~ 2.2 Ga (Bronner and Chauvel, 1979). The Ghallaman sector is separated from Paleoproterozoic basement of Reguibat Shield by Sfariat Belt. The basement gneisses are dated at ~ 3.0 Ga, which was intruded by a ~ 2.9 Ga granite (Schofield et al., 2012). The boundary between Ghallaman and Tiris is not clearly defined due to younger sedimentary cover (Schofield et al., 2012).

Eburnean terranes occupy northeast Mauritania and southwest Algeria, exposed within the Paleoproterozoic Yetti and Eglab massifs (Fig. 13.1B). Yetti massif is characterized by a strongly folded 2.1-Ga volcano-sedimentary series, known as the Lower Reguibat Complex (Peucat et al., 2005) (Fig. 13.2). To the east, Eglab massif is mainly composed of 2.2 Ga foliated volcano-sedimentary rocks (Oued Souss series) and the younger 2.07 Ga syn- and post-orogenic felsic intrusions (Eglab and Aftout series), which are defined as Upper Reguibat Complex (Peucat et al., 2005). Relict oceanic crust of ~ 2.7 Ga in age outcrops in west Eglab massif (Peucat et al., 2005). Paleoproterozoic basement rocks were thrust onto Archean basement of Reguibat Shield at ~ 2.07 Ga along the Sfariat Belt, but only had minimal metamorphic reworking to reactivate biotite (Rocci et al., 1991). Thereafter, Reguibat Shield stabilized and cratonized, and no further history is recorded until Taoudeni deposition in the latest Mesoproterozoic (Rooney et al., 2010).

13.2.3 Anti-Atlas Belt

Paleoproterozoic basement rocks in Anti-Atlas Belt are located in Morocco, bounded by South Atlas Fault in the north and the Tindouf Basin in the south (Fig. 13.1B). The lithostratigraphic framework of Anti-Atlas Belt is well-constructed by Thomas et al. (2004). Anti-Atlas Belt can be subdivided into two parts by the Anti-Atlas Major Fault (AAMF) (Fig. 13.2). Inliers southwest of AAMF contain ~ 2.2 Ga supracrustal schists, gneisses, and migmatites, referred to as Kerdous-Zenaga Complex (Thomas et al., 2004). The metamorphic grade is generally lower toward the west. Younger granitic intrusions with ages around 2.05 Ga are ubiquitous in all inliers west of AAMF (Thomas et al., 2004). In Tagragra de Tata and Zenaga inliers, 2.04 Ga mafic dykes are also observed (Walsh et al., 2002; Kouyaté et al., 2013). The Kerdous-Zenaga Complex and subsequent magmatic suites are correlated with contemporaneous Eburnean events in Reguibat Shield. Following this cratonization, the next stratigraphic event indicates regional extension, as manifested by 1.75 Ga mafic dykes observed in Tagragra d'Akka, Iguerda, and Zenaga inliers (Youbi et al., 2013). The Tasserda-Taghatine Group (quartzite and carbonates) and overlying Oumoula Formation (mainly quartzite) were deposited across several inliers, and that succession is cut by mafic intrusions, with ages of c.1.71 Ga in Igherm inlier (Ikenne et al., 2017) and c.1.65 Ga in Kerdous inlier (Ait Lahna et al., 2020). At ~ 1.41–1.38 Ga and ~ 0.88–0.85 Ga, mafic dykes intruded Bas Draa inlier, and Iguerda and Zenaga inliers, respectively (Fig. 13.2; El Bahat et al., 2013; Söderlund et al., 2013a; Kouyaté et al., 2013), which might be associated with the breakup of supercontinents Nuna/Columbia and Rodinia, respectively. The Tizi n'Tghatine Group, which mainly consists of interbedded siliciclastics and carbonates, are suggested to be late Mesoproterozoic to early-Neoproterozoic in age (Bouougri et al., 2020; Ait Lahna et al., 2020). Extensive early-Neoproterozoic lava basalts and associated dykes (Tachdamt Formation) have been identified across several Anti-Atlas inliers (Fig. 13.2; Álvaro et al., 2014; Ait Lahna et al., 2020). In mid-Neoproterozoic time, obduction and arc accretion occurred along the AAMF, which involved Bou Azzer Group (ophiolites) and Saghro Group (volcanics and turbidites). The siliciclastic-dominated Bleïda Formation found in many inliers is likely mid-late Neoproterozoic in age (Bouougri et al., 2020; Ait Lahna et al., 2020). During the Pan-African orogeny, the Ouarzazate Group, which consists of thick volcano-sedimentary accumulations, extensively covered Anti-Atlas Belt (Thomas et al., 2002, 2004; Blein et al., 2014a,b). Glacial deposits of ~ 590–580 Ma were also found in a few inliers, which could be equivalent to Gaskiers glaciation (Letsch et al., 2018). In early-Cambrian time, thick carbonate rocks (Taroudant Group and correlative successions) were deposited across the Anti-Atlas Belt and northern Tindouf Basin (Bertrand-Sarfati et al., 1991); these deposits include minor volcanogenic components in a likely rift setting (Pouclet et al., 2018).

Compilation of magmatic ages in Man-Leo and Reguibat Shields and Anti-Atlas Belt all show the significant 2.2–2.05-Ga peaks during Eburnean orogeny (Grenholm et al., 2019), which could indicate that the basement of WAC was formed by a series of accretion of Paleoproterozoic rocks to Archean nuclei in Man-Leo and Reguibat Shields; and shortly afterward, Anti-Atlas Belt was accreted to Reguibat Shield at the northern margin. The Taoudeni Basin onlaps both Man-Leo and Reguibat Shields, further demonstrating their connection before 1.1 Ga. Coeval c.1.76–1.75 Ga and c.0.88–0.85 Ga mafic magmatism in Man-Leo Shield and Anti-Atlas Belt provides additional evidence for a united WAC since the Paleoproterozoic (Fig. 13.2).

13.3 Review of paleomagnetic data

Here, all available paleomagnetic data from WAC are compiled on the basis of the Precambrian database PALEOMAGIA (Veikkolainen et al., 2017), the Global Paleomagnetic Database (GPMDB; Pisarevsky, 2005), and the recent Nordic Paleomagnetic Workshops (Evans et al., this issue) (Fig. 13.3; Table 13.1). We try to provide a careful evaluation of these data, as well as pointing out the directions for future paleomagnetic studies in WAC.

The attempt of paleomagnetic studies of Precambrian rocks in WAC was first made in the early 1970s, and then culminated in the late 1970s – 80s. Specifically, in Man-Leo Shield, Piper and Lomax (1973) reported the first dataset from Paleoproterozoic mafic intrusions and a greenstone body in Ghana and Ivory Coast. Later on, Onstott et al. (1984) studied the Paleoproterozoic amphibolite in Liberia and also conducted $^{40}\text{Ar}/^{39}\text{Ar}$ dating of these rocks. A subsequent study by Onstott and Dorbor (1987) presented paleomagnetic and geochronological data from Paleoproterozoic and early Mesoproterozoic lower-amphibolite to granulite facies metamorphic rocks in a nearby area in Liberia. The most recent work in Man-Leo Shield is provided by Nomade et al. (2003), who analyzed the magnetization of mid-Paleoproterozoic granite in Ivory Coast. In Reguibat Shield, Lomax (1975) and Sabaté and Lomax (1975) first documented the paleomagnetism of several late Paleoproterozoic and early Mesoproterozoic mafic intrusions from Yetti-Eglab region. Aifa et al. (2001) mentioned results from 1.9 Ga to 1.4 Ga dyke swarms in Eglab region, but

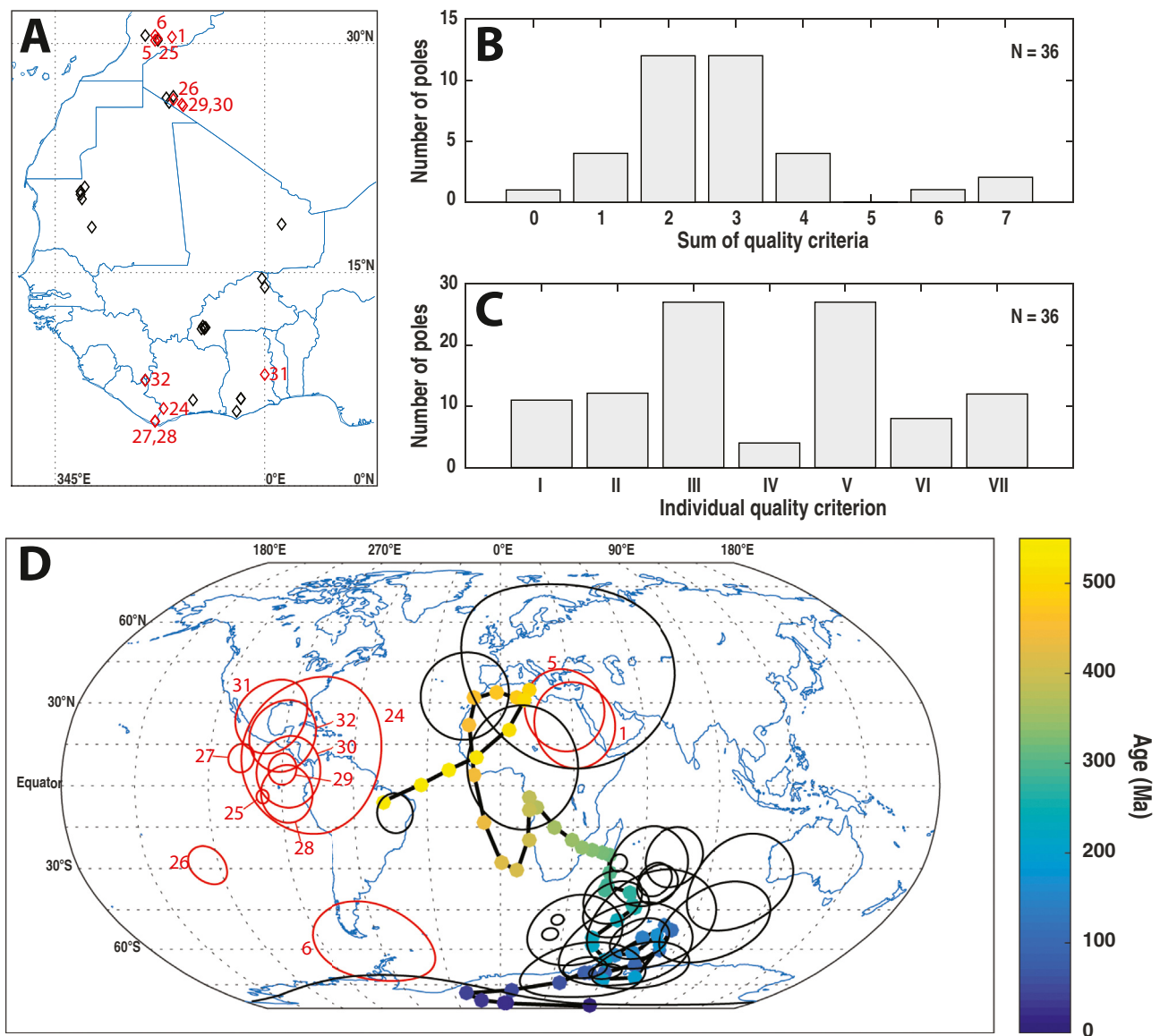


FIGURE 13.3 (A) Distribution of Precambrian paleomagnetic data in West African Craton. Black and red diamonds show paleopoles that resemble and differ from Phanerozoic poles, respectively (B) Sum of quality criteria (Van der Voo, 1990). (C) Individual quality criterion (Van der Voo, 1990). N, total number of paleopoles. (D) Projection of paleopoles with respect to the Phanerozoic apparent polar wander path of West African Craton. Black and red poles represent the ones that overlap and nonoverlap with Phanerozoic poles, respectively. The Phanerozoic running mean polar path of West African Craton is from Torsvik et al. (2012). Pole numbers refer to Table 13.1.

since no data were presented in that paper, we excluded their study from Table 13.1. Sedimentary rocks in the Taoudeni Basin have also been studied. Results from the late-Mesoproterozoic Char and Atar Groups in Mauritania were reported in Morris and Carmichael (1978), Perrin et al. (1988), and Perrin and Prévot (1988). Younger sequences were also sampled by Kent et al. (1984) in Mauritania, and by Boudzoumou et al. (2011) in Burkina Faso, and were assigned to be Cryogenian or Ediacaran based on stratigraphic correlation. The Pan-African Adma diorite (c.616 Ma) in Mali was also studied paleomagnetically by Morel (1981), but technically speaking that unit lies to the east of the craton in the Hoggar Shield. In Anti-Atlas Belt, Martin et al. (1978) discussed the magnetization of Ediacaran sediments in Morocco for the first time. Recently, two new papers have been published regarding the data from Anti-Atlas Belt. Neres et al. (2016) studied several mafic intrusions in Iguerda inlier in Morocco, with two dykes dated to be 885 ± 13 Ma (Kouyaté et al., 2013) and 1746.8 ± 3.7 Ma (Youbi et al., 2013). The most recent study provided data from the Ouazazate and Taroudant Groups in Agadir Melloul, Iguerda, and Bou Azzer inliers in Morocco, for which Ediacaran volcanic deposits and lava flows were analyzed (Robert et al., 2017). In general, paleomagnetic data cover all

TABLE 13.1 Precambrian paleomagnetic data from West African Craton.

#	Formation (Paleopole name)	Slat (°N)	Slon (°E)	Nominal age (Ma)	Met	Rock isotopic age (Ma)	B/N	Dec (°)	Inc (°)	α ₉₅ (°)	k	Plat (°N)	Plon (°E)	A ₉₅ (°)	Reliability criteria												References	
															1	2	3	4	5	6	7	Σ	Paleomagnetism	Geochronology				
1	Djebel Boho volcanics (B2)	30.4	353.4	547	g	541 ± 6	16 ^a /78	87.7	71.9	9.9	14.9	27.3	27.1	14.9	1	1	1	1	1	1	1	7	Robert et al. (2017)	Blein et al. (2014a)				
2	Iguerda mafic dykes (Cluster D4)	30.2	352.4	550	a		2/9 ^a	45.9	-56.7	5.8	79.0	-10.0	316.8	7.2	0	0	1	0	1	0	0	2	Neres et al. (2016)					
3	Azougui Formation (I2)	20.5	346.9	550	a		4 ^a /60	339.0	-17.3	3.1	34.0	-54.1	24.1	2.3	0	1	0	0	1	0	0	2	Perrin and Prévot (1988)					
4	Serie lie de vin sediments (M0)	30.5	351.5	557	a, s		8 ^a /8	124.0	-1.7	20.5	8.2	-29.2	63.3	14.5	0	0	0	0	1	0	0	1	Martin et al. (1978)					
5	Fajjoud and Tadoughast volcanics (B1)	30.2	352.2	562	g	556 ± 5	20 ^a /99	93.7	67.0	10.5	10.7	21.9	31.0	15.6	1	1	1	1	1	1	1	7	Robert et al. (2017)	Blein et al. (2014b)				
6	Adrar-n-takoucht Volcanics (C)	30.5	352.2	571	g	571 ± 6	9 ^a /53	27.7	18.0	19.4	8.0	-57.6	295.6	15.7	1	0	1	1	1	1	1	6	Robert et al. (2017)	Blein et al. (2014b)				
7	Mejeria red sandstone	18.1	347.7	582	a, s		4 ^a /17	137.1	14.4	13.2	49.0	-40.5	49.9	9.7	0	0	0	0	1	0	0	1	Kent et al. (1984)					
8	Oujeft Group (CO8 + CO10)	20.0	347.0	600	a, e	595 ± 45	2/14 ^a	127.0	23.0	4.0	89.0	-28.0	50.0	3.1	0	0	1	0	1	0	0	2	Morris and Carmichael (1978)	Clauer et al. (1982)				
9	Adma diorite	18.3	1.2	616	g	616 ± 11	13 ^a /62	317.3	78.6	9.0	22.0	32.5	344.7	15.9	1	1	1	0	1	0	0	4	Morel (1981)	per. comm. w/Ducrot, Lancelot, and Renaud				
10	Tin Akof carbonates (TA)	14.0	0.0	618	a		1/7 ^a	332.1	-0.1	16.6	14.2	-58.8	64.6	16.6	0	0	1	1	1	1	0	4	Boudzoumou et al. (2011)					
11	Tin Dioulaf cap carbonates (TD)	14.6	359.8	635	corr.		1/10 ^a	130.5	16.2	7.0	48.5	-35.8	68.5	7.2	1	0	1	0	1	0	0	3	Boudzoumou et al. (2011)	Boudzoumou et al. (2011)				
12	Guinguette sediments (G)	11.1	355.7	715	a		1/8 ^a	322.6	-11.5	18.5	10.0	-49.1	71.7	18.8	0	0	1	0	1	0	0	2	Boudzoumou et al. (2011)					
13	Banarodougou sediments (B)	11.3	355.6	715	a		1/15 ^a	345.0	-10.8	4.4	75.5	-67.6	46.9	4.5	0	0	1	0	1	0	0	2	Boudzoumou et al. (2011)					
14	Tiara sediments (Tr)	11.2	355.8	715	a		1/18 ^a	343.1	-2.2	3.2	107.0	-69.2	59.1	3.2	0	0	1	0	1	0	0	2	Boudzoumou et al. (2011)					
15	Digouéra sediments (D)	11.1	355.5	715	a		1/13 ^a	346.0	-8.6	11.9	13.0	-69.2	47.3	12.0	0	0	1	0	1	1	0	3	Boudzoumou et al. (2011)					
16	Toussiana sediments (T)	11.2	355.7	775	a	775 ± 52	1/9 ^a	300.2	-26.1	10.1	27.0	-25.9	73.4	10.9	0	0	1	0	1	1	0	3	Boudzoumou et al. (2011)					
17	Atar group, Unit I9	20.8	347.2	1100	l	1107 ± 12, 1109 ± 22, 1105 ± 37	2/15 ^a	125.2	-6.8	8.6	20.0	-34.0	66.9	6.1	0	0	1	0	1	1	0	3	Perrin et al. (1988)	Rooney et al. (2010)				
18	Char Group (I2)	20.3	346.9	1150	s, e	998 ± 34	5 ^a /25	336.0	37.0	11.0	36.0	-67.0	78.0	9.9	0	1	1	0	1	0	0	3	Morris and Carmichael (1978)	Clauer et al. (1982); Rooney et al. (2010)				
19	Char group, Unit I2	20.5	346.9	1150	s, e	998 ± 34	4 ^a /52	334.4	-23.6	2.5	63.0	-48.7	26.6	1.9	0	1	1	0	1	0	0	3	Perrin et al. (1988)	Clauer et al. (1982); Rooney et al. (2010)				

(Continued)

TABLE 13.1 (Continued)

#	Formation (Paleopole name)	Slat (°N)	Slon (°E)	Nominal age (Ma)	Met	Rock isotopic age (Ma)	B/N	Dec (°)	Inc (°)	α ₉₅ (°)	k	Plat (°N)	Plon (°E)	A ₉₅ (°)	Reliability criteria												References	
															1	2	3	4	5	6	7	Σ	Paleomagnetism	Geochronology				
20	Nimba-Harper metamorphic rocks (NiB)	7.5	351.5	1480	c	1450 ± 10, 1510 ± 10	4 ^a /22	140.2	26.7	13.0	51.0	-45.0	52.8	10.4	1	0	1	0	0	1	0	3	Onstott and Dorbor (1987)	Onstott et al. (1984); Onstott and Dorbor (1987)				
21	Shahra dykes	26.6	353.0	1500	s		7 ^a /34	55.1	71.1	21.5	8.9	41.2	31.0	35.0	0	0	0	0	1	0	0	1	Sabaté and Lomax (1975)					
22	Quartz diorite	26.3	353.2	1500	s		3 ^a /18	139.5	77.0	12.5	97.9	6.6	9.1	22.5	0	0	0	0	0	0	0	0	Sabaté and Lomax (1975)					
23	Dykes intruding Guelb el Hadid Series	26.7	353.5	1500	s		9 ^a /53	296.6	50.6	16.0	11.3	-35.1	104.9	17.7	0	1	0	0	1	0	0	2	Sabaté and Lomax (1975)					
24	Ivory Coast dolerite intrusions	5.5	352.8	1730	d	1730 ± 170	1/10 ^a	281.0	14.0	39.0	2.5	11.0	282.0	28.5	0	0	0	0	1	0	1	1	Piper and Lomax (1973)	Knopf (1967)				
25	Iguerda mafic dyke (Cluster D1)	30.3	352.4	1747	g	1747 ± 4	1/22 ^a	86.7	4.6	3.5	81.0	-4.0	262.1	2.5	1	0	1	0	1	0	1	4	Neres et al. (2016)	Youbi et al. (2013)				
26	Aftout gabbro	26.5	353.5	1869	a	1869 ± 50	5 ^a /26	72.4	55.5	5.7	179.0	-28.8	235.1	6.9	0	1	0	0	0	0	1	2	Sabaté and Lomax (1975)					
27	Harper amphibolite (HaA1)	4.6	352.2	1950	e, c	1983 ± 34, 1964 ± 2, 1894 ± 2	3/11 ^a	274.0	-14.0	7.0	39.0	10.0	253.0	5.2	1	0	1	0	0	0	1	3	Onstott et al. (1984)	Onstott et al. (1984)				
28	Harper amphibolite (HaA2)	4.6	352.2	1950	e, c	1983 ± 34, 1964 ± 2, 1894 ± 2	1/3 ^a	89.0	-19.0	14.0	76.0	-3.0	272.0	10.5	1	0	1	0	0	0	1	3	Onstott et al. (1984)	Onstott et al. (1984)				
29	Aftout plutons (E)	26.2	354.1	1966	s		6 ^a /40	92.8	-15.5	7.8	75.1	6.0	270.1	5.7	0	1	1	0	0	0	1	3	Lomax (1975)					
30	Aftout plutons (C)	26.1	354.2	1966	s		4 ^a /24	90.8	-19.2	17.2	29.4	5.0	272.6	13.0	0	0	1	0	0	0	1	2	Lomax (1975)					
31	Nimba-Harper metamorphic rocks (NiA)	7.5	351.5	2050	c	2050 ± 6	6 ^a /38	286.6	18.4	13.0	28.0	18.0	269.0	13.0	1	1	1	0	0	0	1	4	Onstott and Dorbor (1987)	Onstott et al. (1984); Onstott and Dorbor (1987)				
32	Ferke batholith (IC2)	7.9	0.0	2094	j	2094 ± 6	9 ^a /40	296.0	3.0	18.9	8.4	25.0	263.0	13.4	0	0	1	0	0	0	1	2	Nomade et al. (2003)	Doumbia et al. (1998)				
33	Obuasi dolerite dyke	6.2	358.3	2100	s		1/10 ^a	328.0	-11.0	11.0	20.9	-56.0	69.0	7.9	0	0	1	0	1	0	0	2	Piper and Lomax (1973)					
34	Tarkwa dolerite intrusions	5.3	358.0	2100	s		5 ^a /29	156.0	40.0	14.0	32.0	-53.0	36.0	13.5	0	1	1	0	1	0	0	3	Piper and Lomax (1973)					
35	Ivory Coast TTG plutons (IC1)	6.1	354.9	2100	s, a		6 ^a /24	8.4	3.7	16.1	18.3	-82.0	292.0	11.4	0	1	1	0	1	0	0	3	Nomade et al. (2003)					
36	Obuasi greenstone body	6.2	358.3	2200	g	~2200	5 ^a /74	320.0	26.0	19.0	20.7	-50.0	102.0	15.3	1	0	0	0	1	0	0	2	Piper and Lomax (1973)	per. comm. w/P.G. Coomer				

^alevel of mean calculation, Dec, declination; Inc, inclination; Plat/Plon, paleopole latitude/longitude; α_{95} , radius of 95% confidence cone of the mean direction; k, precision parameter; A_{95} , radius of 95% confidence cone of the paleomagnetic pole. Reliability criteria follow Van der Voo (1990). Bold font highlights paleopoles that nonoverlap with the Phanerozoic apparent polar wander path of West African Craton.

Note: Slat/Slon = site geographic latitude/longitude, Met = magnetization age determination method following PALEOMAGIA Database (a, geological/APWP; c, ⁴⁰Ar/³⁹Ar; corr., correlation; d, K-Ar; e, Rb-Sr; g, U-Pb; j, Pb-Pb; l, Re-Os; s, stratigraphic), B/N = number of sites/samples.

cratonic parts in WAC (Fig. 13.3A). However, as of now, fewer than 20 papers have been published, with only four in this century, which makes WAC the least studied region paleomagnetically compared to other cratons.

We evaluate Precambrian paleomagnetic data in WAC following the quality criteria proposed by Van der Voo (1990) (Fig. 13.3B–D). The current WAC dataset faces five major problems. First and foremost is the age constraint on the rocks, as well as the age of magnetization. It is noted that the ages of the rocks in the studies published in the 1970s and 1980s usually are constrained by stratigraphic correlation, or merely by geological cross-cutting relationships. Even though some studies gave radiometric ages, these ages have issues in either the precision or the accuracy or both. Another wrinkle to this problem is that metamorphic rocks in some of these studies could have quite complicated thermal histories, therefore the age of the magnetization could be very different from the age of the rock. Second, most of the old works can only be counted as reconnaissance studies since they often applied a spotty sampling of rocks with different ages across a large area, without fully investigating a particular rock unit. For example, two-thirds of the paleopoles fail to satisfy the second criterion of Van der Voo (1990) (Fig. 13.3C), which means they do not have sufficient number of samples in order to yield any robust paleopole. Nearly 30% of the paleopoles were calculated from data containing only one paleomagnetic sampling site (Fig. 13.3C), which is inadequate for statistical purposes, and the secular variation of ancient geomagnetic field cannot be averaged out. The third problem concerns the demagnetization techniques. Since modern demagnetization methods and/or high-precision rock magnetometers were either not available or just becoming established four decades ago, most samples in the earliest literature were not adequately demagnetized in high temperatures or alternating fields. It is possible that data from those samples are partly contaminated by overprints to various degrees. Fortunately, this problem becomes much less relevant when dealing with more recent publications. Fourth, Precambrian WAC data rarely have field tests to determine if the magnetization is primary. The last problem is that some older paleopoles show resemblance to younger poles, which indicates the suspicion of remagnetization. When we plotted WAC dataset against the Phanerozoic apparent polar wander path (APWP) of WAC, we found that many paleopoles overlap with the late-Paleozoic and early-Mesozoic segment of the path (Fig. 13.3D). A resemblance to younger poles could be just coincidental, if considering a long timescale of hundreds of million years or more (Pivarunas et al., 2018). However, we believe that the resemblances shown in the WAC dataset are likely evidence for remagnetizations. In Table 13.1, we can see that paleopoles similar to late-Carboniferous segment have south-seeking declinations, which should be predominant during the Kiaman reverse superchron (Torsvik et al., 2012). This remagnetization is likely related to the Hercynian/Variscan orogeny when Gondwana collided with Laurussia to build the Pangea supercontinent. Paleopoles similar to the Permian and early-Mesozoic segment have very shallow inclinations, when WAC occupied low-latitude areas (Torsvik et al., 2012). But we should point out that this remagnetization issue should not discourage future research in WAC, because some studied rocks could still preserve the primary information, which was not able to be isolated in older studies due to inadequate demagnetization and low-precision measurement.

In fact, promising data still exist in WAC dataset. Five paleopoles of Ediacaran age fulfill at least four quality criteria of Van der Voo (1990), especially the three yielded by the Ouazazate and Taroudant volcanics (Robert et al., 2017). With a positive fold test and a positive intraformational conglomerate test, these three paleopoles are believed to be robust (Robert et al., 2017), and should contribute to a better understanding of the complexity of Ediacaran geomagnetic field and the paleogeography of WAC. It is also noticed that there is a paleopole clustering around Gulf of Mexico – Caribbean Sea, for which the ages range from 2050 Ma to 1750 Ma (Fig. 13.3D). This paleopole clustering might have potential in providing a useful Orosirian pole for WAC, but awaits more paleomagnetic and geochronological investigations.

Overall, huge gaps remain in the paleomagnetic record of Precambrian rocks in WAC (Evans et al., *this issue*), hindering paleogeographic reconstruction in deep time. Future studies should consider all major problems in WAC dataset, by taking advantage of newly published high-precision ages (e.g., Rooney et al., 2010) and modern paleomagnetic and rock magnetic routines. We also suggest mafic dyke swarms originated from the large igneous provinces (LIPs) as research priorities since they are reliable recorders of magnetization, and can be precisely dated at the same time. The geometry of dyke swarms can be compared with coeval swarms in other cratons and restored back for reconstructing the LIPs, and at the same time the paleogeography of cratons. In addition, baked-contact tests can be conducted in paleomagnetic studies of dyke swarms. Previous studies have witnessed many successful cases of using mafic dykes to generate robust paleopoles for Precambrian APWPs (e.g., Buchan et al., 1994). The study of Neres et al. (2016), while itself insufficiently limited to a small number of sampled sites, points the way toward further research on Proterozoic mafic dykes within the Anti-Atlas Belt.

13.4 LIP records in West African Craton

In view of the significance of LIPs in paleogeographic studies, we summarize all precisely dated Precambrian mafic dykes and sills in WAC (Table 13.2), together with the available information about their geometry and geochemistry and suggested tectonic settings. Numerous mafic intrusions (mainly dykes) are ubiquitous in WAC and show clear lineation on satellite images and strong magnetic anomalies. A recent contribution of Jessell et al. (2015) mapped mafic dykes in WAC comprehensively. This new map is mainly based on aeromagnetic data and incorporates all previous publications. It is shown that dykes densely intruded all basement rocks and old basins in WAC (Fig. 13.4). Areas devoid of dykes are usually covered by younger strata so there is no exposure. Varying orientations and cross-cutting relationships of dykes indicate that intrusions should have various ages. In the Anti-Atlas Belt, a total of six swarms have been dated. The oldest 2.04 Ga Tagragra de Tata swarm is defined by two ENE-striking dykes in Tagragra de Tata and Zenaga inliers (Walsh et al., 2002; Kouyaté et al., 2013). The origin of this swarm is proposed to be post-Eburnean extension because the 2.04-Ga dyke postdates all Eburnean-granite in Tagragra de Tata inlier (Walsh et al., 2002). However, Kouyaté et al. (2013) suggest that the granite and mafic dykes are the products of a bimodal magmatic suite, which could be related to the WAC – North Atlantic Craton separation at 2.04 Ga. The 1.75 Ga Tagragra d'Akka swarm has been found in at least four inliers (Tagragra d'Akka, Tafeltast-Kerdous, Zenaga, Iguerda) with a consistent NW strike (Youbi et al., 2013). A c.1.71-Ga sill in Igherm inlier also likely belongs to this event (Ikenne et al., 2017). Titanium/vanadium ratios of the 1.75 Ga dykes show a continental flood basalt or a mid-ocean ridge basalt signature, which indicates a rifting context (Youbi et al., 2013). The 1.65-Ga Zenaga event is named after two sills and a NE-striking dyke in Zenaga inlier (Kouyaté et al., 2013). The 1.41–1.38 Ga Bas Drâa swarm is determined by two dated dykes in Bas Drâa inlier (El Bahat et al., 2013; Söderlund et al., 2013a). Geochemistry (Th/Yb vs Ta/Yb, La/Sm vs Zr/Nb) shows a within-plate volcanic zone signature (El Bahat et al., 2013). The 1.41–1.38 Ga dykes are similar in age to the breakup of the supercontinent Nuna/Columbia; therefore, they are suggested to possibly originate from a hotspot mantle source beneath WAC during rifting (El Bahat et al., 2013). The 0.88–0.85-Ga Iguerda-Taïfast swarm is based on two dated NE-striking dykes in Iguerda-Taïfast inlier (Kouyaté et al., 2013), and they could be associated with the rifting event within the supercontinent Rodinia. A 611-Ma dyke in Bas Drâa inlier is dated by El Bahat et al. (2017) and could belong to a swarm that has not been recognized to its full extent. Southward, the Reguibat Shield has the two oldest dykes thus found in WAC: the 2.73-Ga NE-striking Ahmeyim Great dyke (Tait et al., 2013) and the 2.69-Ga NW-striking Aousserd-Tichla dyke (Söderlund et al., 2013b). The relationship between these two dykes is not clear, and they might represent two separate events. Geochemical analysis points to a boninitic provenance of the 2.73-Ga Ahmeyim Great dyke, but it is suggested to be a signal of crustal contamination and the dyke is unlikely related to subduction (Tait et al., 2013). In Man-Leo Shield, Baratoux et al. (2019) recently published a series of new U-Pb baddeleyite ages of dykes that have been previously dated by K-Ar or Ar-Ar methods. In total, there are six swarms, namely the 1.79-Ga Libiri swarm, the 1.76-Ga Kédougou swarm, the 1.57-Ga Korsimoro swarm, the 1.52-Ga Essakane swarm, the 0.91-Ga Oda swarm, and the 0.85-Ga Manso swarm (Baratoux et al., 2019). Across the craton, the 1.75-Ga Tagragra d'Akka swarm in Anti-Atlas Belt and the 1.76-Ga Kédougou swarm in Man-Leo Shield could be related to the same event because of the matching ages. In addition, the 0.88–0.85-Ga Iguerda-Taïfast swarm in Anti-Atlas Belt and the 0.85-Ga Manso swarm could also originate from a single event. With the rapid growth of geochronological data regarding the mafic intrusions in WAC, we expect more swarms to be identified and the magmatic history of WAC to be better constrained.

Magmatic barcode comparisons are made between WAC and its potential neighboring cratons in Proterozoic supercontinents, for example, Amazonia, Baltica, Laurentia, and Siberia (Fig. 13.5). During certain time intervals, several cratons share similar-aged magmatism. For instance, in WAC, the 2.04-Ga swarm is coeval with the Korak and Kangâmuit-MD3 swarms in North Atlantic Craton (Nilsson et al., 2013); the 1.79-Ga swarm matches the Avanavero swarm in Amazonia (Reis et al., 2013), the Xiong'er swarm in North China (Peng, 2010), the Tomashgorod swarm in Baltica (Bogdanova et al., 2013), and the Florida swarm in Río de la Plata (Teixeira et al., 2013); the 1.75-Ga swarm could be correlated with the Cleaver-Hadley Bay-Nueltin swarm in Laurentia (Ernst et al., 2013), the Espinhaço swarm in São Francisco (Danderfer et al., 2009), and the Chaya swarm in Siberia (Gladkochub et al., 2010); the 1.65-Ga swarm is comparable with the Khibelen swarm in Siberia (Gladkochub et al., 2007) and the Häme swarm in Baltica (Salminen et al., 2017); the 1.52-Ga swarm is likely associated with the Käyser swarm in Amazonia (Baratoux et al., 2019); the 1.41–1.38-Ga swarm is of the same age as the Mashak swarm in Baltica (Puchkov et al., 2013), the Chieress swarm in Siberia (Ernst et al., 2000), the Nova Lacerda swarm in Amazonia (Teixeira et al., 2016), the Kunene event in Congo (McCourt et al., 2013), and the Hart River-Salmon

TABLE 13.2 Geochronology of Precambrian mafic intrusions in West African Craton.

Location	Swarm	Latitude (°N)	Longitude (°E)	Sample name	Dyke trend	Age (Ma)	Method	References
Anti-Atlas, Tagragra de Tata	Tagragra de Tata	29.9058	352.0117	41801	N105°	2039.5 ± 6.3	SHRIMP on zircon	Walsh et al. (2002)
Anti-Atlas, Zenaga	Tagragra de Tata	30.3713	352.6913	DZ49	N100°	2040 ± 2	ID-TIMS U-Pb on zircon	Kouyaté et al. (2013)
Anti-Atlas, Tagragra d'Akka	Tagragra d'Akka	29.3728	351.1764	DAK10	N125°	1758 ± 5	ID-TIMS U-Pb on zircon	Youbi et al. (2013)
Anti-Atlas, Tafeltast-Kerdous	Tagragra d'Akka	29.4077	351.0089	AT14	N170°	1741 ± 10	ID-TIMS U-Pb on baddeleyite	Youbi et al. (2013)
Anti-Atlas, Iguerda-Taïfast	Tagragra d'Akka	30.2602	352.4015	DIGI064	N135°	1746.8 ± 3.7	ID-TIMS U-Pb on baddeleyite	Youbi et al. (2013)
Anti-Atlas, Zenaga	Tagragra d'Akka	30.5221	352.8222	DZ67	N125°	1734 ± 5	ID-TIMS U-Pb on baddeleyite	Youbi et al. (2013)
Anti-Atlas, Igherm	Tagragra d'Akka	29.9983	351.3500	IM3	(sill)	1706 ± 7	ID-TIMS U-Pb on baddeleyite	Ikenne et al. (2017)
Anti-Atlas, Zenaga	Zenaga	30.5841	352.8218	DZ40	(sill)	1656 ± 9	ID-TIMS U-Pb on baddeleyite	Kouyaté et al. (2013)
Anti-Atlas, Zenaga	Zenaga	30.5828	352.8173	DZ36	(sill)	1641 ± 3	ID-TIMS U-Pb on baddeleyite	Kouyaté et al. (2013)
Anti-Atlas, Agadir Melloul	Zenaga	30.2460	352.1994	AMHA069	N45°	1654 ± 16	ID-TIMS U-Pb on baddeleyite	Kouyaté et al. (2013)
Anti-Atlas, Bas Drâa	Bas Drâa	28.4818	349.2164	BD21	N40°	1384 ± 6	ID-TIMS U-Pb on baddeleyite	El Bahat et al. (2013)
						1416 ± 7	LA-ICP-MS U-Pb on baddeleyite	Söderlund et al. (2013a)
Anti-Atlas, Bas Drâa	Bas Drâa	28.4603	349.4232	BD12	N135°	1381 ± 8	ID-TIMS U-Pb on baddeleyite	El Bahat et al. (2013)
Anti-Atlas, Iguerda-Taïfast	Iguerda-Taïfast	30.1905	352.3044	DAS094	N30°	885 ± 13	ID-TIMS U-Pb on zircon	Kouyaté et al. (2013)
Anti-Atlas, Zenaga	Iguerda-Taïfast	30.5065	352.7611	DZ2	N30°	857 ± 11	ID-TIMS U-Pb on baddeleyite	Kouyaté et al. (2013)
Anti-Atlas, Bas Drâa	—	28.4573	349.4225	D16	N75°	611.6 ± 1.3	⁴⁰ Ar/ ³⁹ Ar on hornblende	El Bahat et al. (2017)
Reguibat	Ahmeyim	20.7163	345.4753	GTD-8	N30°	2732.6 ± 1.5	ID-TIMS U-Pb on baddeleyite	Tait et al. (2013)
Reguibat	Aousserd-Tichla	22.3984	345.6255	CMS-25	N125°	2688 ± 3	ID-TIMS U-Pb on baddeleyite	Söderlund et al. (2013b)
Man-Leo	Libiri	13.3980	1.2548	DO109	N5–10°	1791 ± 3	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Kédougou	12.6827	347.8080	BK031	N30–40°	1764 ± 4	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Korsimoro	12.8860	358.9295	DO112	N100°	1575 ± 5	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Essakane	14.4271	359.9812	DO115	N130°	1519 ± 2	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Essakane	11.0774	359.5137	YU14	N130°	1525 ± 3	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Essakane	13.4974	359.6712	DO09	N130°	1523 ± 2	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Essakane	9.9983	357.5661	BN29	N130°	1521 ± 2	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Essakane (Sambarabougou)	13.1036	348.0760	BK001	N90°	1521 ± 3	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Oda	5.4338	358.8423	MG-8	N70°	915 ± 7	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)
Man-Leo	Manso	5.3083	358.1293	GH370	N175°	867 ± 16	ID-TIMS U-Pb on baddeleyite	Baratoux et al. (2019)

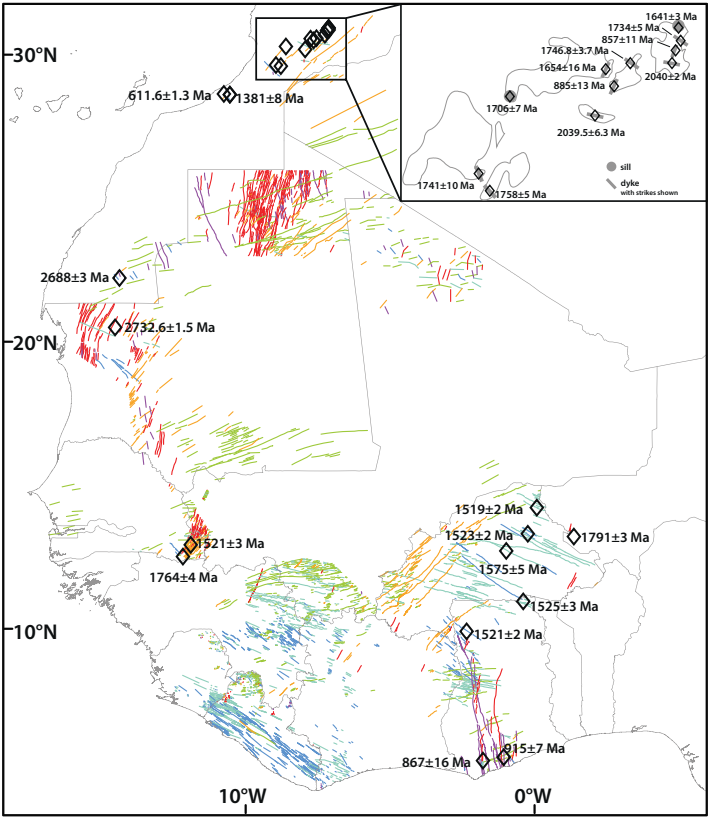


FIGURE 13.4 Mafic dyke distribution across West African Craton. Black diamonds mark the locations of the dated dykes (Jessell et al., 2015). Modified from Jessell, M., Santoul, J., Baratoux, L., Youbi, N., Ernst, R.E., Metelka, V., et al., 2015. An updated map of West African mafic dykes. *Journal of African Earth Sciences* 112, 440–450.

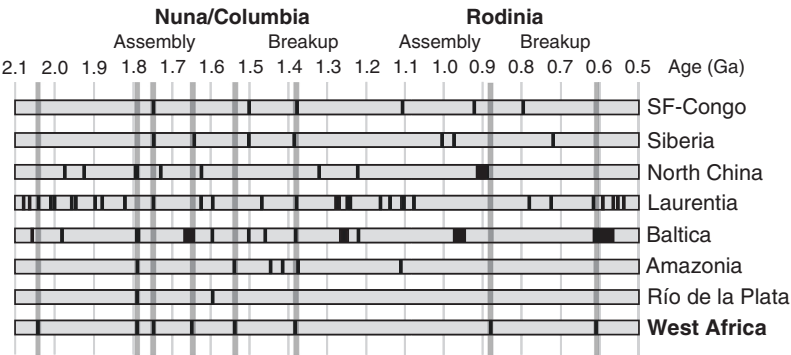


FIGURE 13.5 LIP barcodes of Proterozoic cratons. Thin lines represent single pulses and wide boxes indicate prolonged pulses for a certain event. Note potential correlations between West African Craton and other cratons are shown by shaded bars (correlation window size = 15 Myr). SF, São Francisco (Ernst et al., 2013). Source: Modified from Ernst, R.E., Bleeker, W., Söderlund, U., Kerr, A.C., 2013. Large igneous provinces and supercontinents: toward completing the plate tectonic revolution. *Lithos* 174, 1–14.

Arch and Midsommerso-Zig Zag Dal swarms in Laurentia (Ernst et al., 2013); the 0.88-Ga swarm may be linked with the late stage of the Dashigou swarm in North China (Peng et al., 2011); and the 0.61-Ga swarm could be connected to the early stage of the Central Iapetus Magmatic Province in Laurentia and Baltica (Ernst and Bell, 2010). Although it seems that penecontemporaneous swarms can be traced across several cratons, we emphasize that these temporal matches should be treated as indicative rather than diagnostic. Besides, the tectonic backgrounds of these swarms are still debated, for example, whether they are LIPs events during continental rifting or regional extensions. Nevertheless, paleomagnetism of these swarms can provide an independent test to constrain the relative positions between WAC and possible neighboring cratons in future studies.

13.5 Paleoclimate indicators

Geological records preserve information about Earth’s climate in the past. Commonly used paleoclimate indicators can be subdivided into three categories: lithological, geochemical, and paleontological. Some of the indicators

can be paleogeographically indicative as well. Here, we summarize available paleoclimate indicators in WAC, plotting against the reconstruction of WAC's paleolatitude based on the paleomagnetic data. It is worthwhile to mention the uncertainties of paleolatitudinal reconstructions before late Neoproterozoic time. First, the geomagnetic polarity is not constrained, so WAC can be placed in either hemisphere. Second, because the quality of paleomagnetic data is low, the inferred paleolatitude is less robust compared to the late Neoproterozoic.

Two important Paleoproterozoic iron formations are found in WAC: the c.2.2 Ga Ijil Group in Mauritania, Reguibat Shield (Bronner and Chauvel, 1979), and the c.2.3 Ga Nimba Itabirite in Liberia, Man-Leo Shield (Berge, 1974). The origin of Paleoproterozoic iron formation is highly debated, but it is suggested that the iron is either sourced from hydrothermal systems or continental inputs, and was deposited due to the enhanced oxygen availability in the ocean (Konhauser et al., 2017). During the deposition of these two iron formations, WAC was likely located at low- to mid-latitudes (Fig. 13.6). In the latest Mesoproterozoic, the Char and Atar Groups of the Taoudeni Basin are characterized by carbonates, with columnar stromatolites and evaporites (Bertrand-Sarfati et al., 1991). Paleomagnetic inclinations of Char and Atar carbonates, if primary, imply a low-latitudinal position of WAC (Perrin et al., 1988), which would be consistent with the depositional environment of carbonates and evaporites. During the Ediacaran, large latitudinal shift of WAC is indicated by paleomagnetic data, which could be caused by a rapid true polar wander event (Robert et al., 2017). Marinoan-equivalent diamictites are found in the Assabet el Hassiane Group of the Taoudeni Basin (Bertrand-Sarfati et al., 1991). In addition, possible Gaskiers-equivalent diamictites are observed in the Taoudeni Basin and the Anti-Atlas Belt (Deynoux, 1985; Letsch et al., 2018; Fig. 13.6). In the Anti-Atlas Belt, massive carbonates were deposited in Taroudant Group (Bertrand-Sarfati et al., 1991), where Ediacara-biota have also been identified (Fig. 13.6; Letsch et al., 2019).

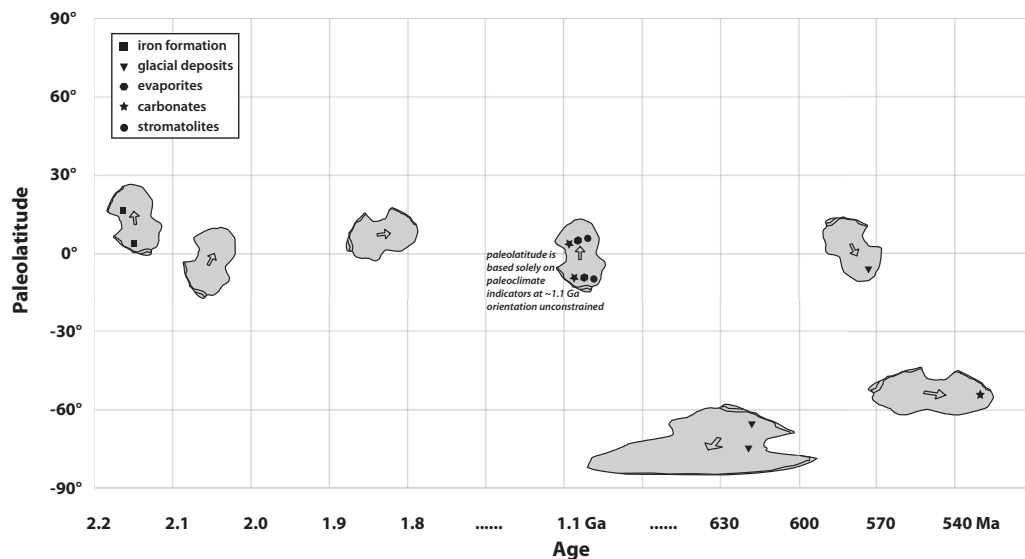


FIGURE 13.6 Paleoclimate indicators. Paleolongitude is arbitrary. Due to the magnetic polarity flexibility, West African Craton can be placed in either hemisphere before late Neoproterozoic time. Paleolatitude of West African Craton at ~ 1.1 Ga is constrained solely by paleoclimate indicators. Arrows point to present-day north. Note that the age scale is not linear.

13.6 Precambrian paleogeography of West African Craton

Discussions regarding the Precambrian paleogeography of WAC are reviewed here, ranging from its possible positions in Atlantica supercraton, and Nuna/Columbia and Rodinia supercontinents, to the final amalgamation within Gondwana. Both paleomagnetic and geological constraints are considered in reconstructions.

13.6.1 Paleogeographic connection between West African Craton and Amazonia

Many paleogeographic reconstructions treat WAC and Amazonia as a long-lived conjoined entity since Paleoproterozoic time, by linking the southern margin of WAC to the Guiana Shield of Amazonia, until they

were finally separated by the opening of the Atlantic Ocean in the Mesozoic. This idea was first proposed by [Onstott and Hargraves \(1981\)](#), who observed a general overlapping of paleopoles from the two cratons between 2.1 Ga and 1.5 Ga. They also suggested that the lineaments of the Guri Belt (fault zone separating the Archean Imataca domain and the Paleoproterozoic Maroni-Itacaiúnas domain) in Guiana Shield and the Sassandra Belt (fault zone separating the Archean Kénéma-Man domain and the Paleoproterozoic Baoulé-Mossi domain) in Man-Leo Shield should be aligned as a piercing point ([Fig. 13.7A](#)) that was subsequently displaced by more than 1000 km through dextral transcurrent faulting during Pan-African orogeny ([Onstott and Hargraves, 1981](#)). This reconstruction would also align the general trends of the Paleoproterozoic Eburnean and Transamazonian orogenic belts in WAC and Amazonia, respectively.

[Nomade et al. \(2003\)](#) attempted to test this direct WAC – Amazonia connection by comparing the APWPs of both cratons, and suggested that such a connection is paleomagnetically permissive. Recently, new paleomagnetic data from Guiana Shield allows a better-defined APWP to be constructed. On that basis, [Bispo-Santos et al. \(2014\)](#) found that the APWPs of the two cratons begin to match around 1980–1960 Ma ([Fig. 13.7B](#)), and WAC and Amazonia might be linked at that time. It should be pointed out that c.2.0-Ga paleopoles from WAC need more rigorous reliability tests and the precision of their ages should be largely improved. Also, numerous dyke swarms observed in Man-Leo and Guiana Shields seem to indicate that continental rifting occurred several times since 1.9 Ga; if so, why would the geological and paleomagnetic datasets persist in their similarities? Nonetheless, the long-lived connection model requires further tests by developing more comprehensive younger-Proterozoic APWPs from both cratons.

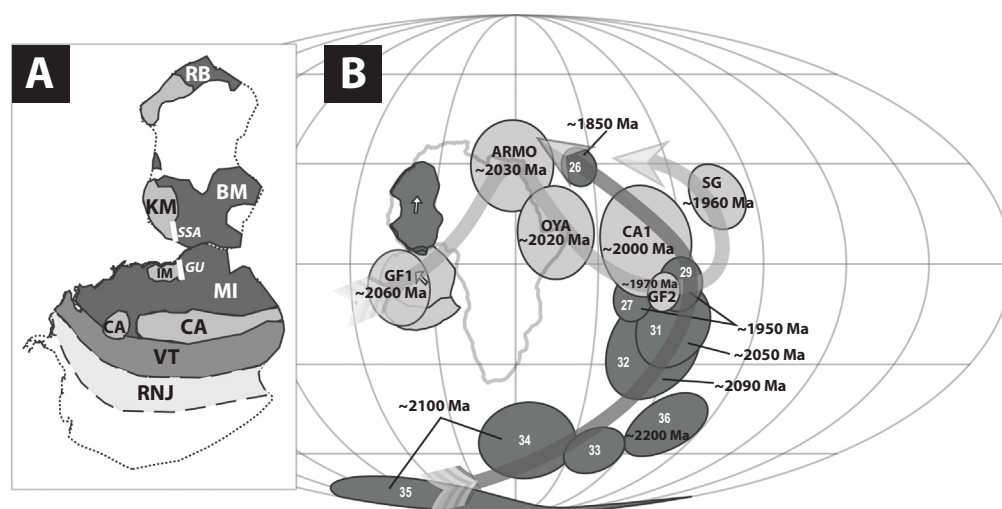


FIGURE 13.7 (A) Paleogeographic connection between West African Craton and Guiana Shield in mid-Paleoproterozoic time. RNJ, Rio Negro-Juruena domain; VT, Ventuari-Tapajós domain; CA, Central Amazonia domain; MI, Maroni-Itacaiúnas domain; IM, Imataca domain; GU, Guri Belt; BM, Baoulé-Mossi domain; KM, Kénéma-Man domain; RB, Reguibat Shield; SSA, Sassandra Belt ([Bispo-Santos et al., 2014](#)). (B) Orosirian apparent polar wander paths of West African Craton and Amazonia in present West Africa reference frame. Euler rotation of Amazonia with respect to absolute reference: 43.3°N, 330.5°E, 71.5°. Paleopoles of West African Craton are numbered as in [Table 13.1](#). Paleopoles of Amazonia are from [Bispo-Santos et al. \(2014\)](#). White arrows show present-day north. Modified from [Bispo-Santos, F., D'Agrella-Filho, M.S., Janikian, L., Reis, N.J., Trindade, R.I., Reis, M.A.A., 2014. Towards Columbia: Paleomagnetism of 1980–1960 Ma Surumu volcanic rocks, Northern Amazonian Craton. Precambrian Research 244, 123–138.](#)

13.6.2 West African Craton in Precambrian supercontinents

It is still highly uncertain whether there ever existed a supercontinent (a.k.a. Kenorland) or a few supercratons in the Neoproterozoic and the earliest Paleoproterozoic. But some ideas have been proposed regarding the configuration of the possible supercratons ([Rogers, 1996](#); [Bleeker, 2003](#)). Atlantica supercraton ([Rogers, 1996](#)) is proposed to be the continental landmass formed at ~ 2 Ga by the collision of WAC, São Francisco-Congo, Amazonia, and Río de la Plata, with their relative positions being similar to those within Pangea. The geological underpinning of Atlantica supercraton is a supposedly correlative ~ 2-Ga fluvio-deltaic sedimentary succession across all of the constituent cratons ([Ledru et al., 1994](#)). [Rogers \(1996\)](#) suggested that Atlantica stayed as a coherent entity since its formation and only experienced intracratonic orogenic activities until the final destruction by the opening of

Atlantic Ocean in the Mesozoic. Regardless of the existing views against the longevity of Atlantica (e.g., D'Agrella-Filho et al., 2011; Evans, 2013), its configuration is also challenged by paleomagnetic data (Rapalini et al., 2015; Franceschinis et al., 2019). For example, by matching the 2.1–1.9-Ga paleopoles of Río de la Plata, Guiana, WAC, and São Francisco – Congo, Franceschinis et al. (2019) suggested that Río de la Plata should sit between Guiana and São Francisco – Congo before 2.0 Ga, and later southeast WAC collided with southeast Río de la Plata to form so-called “unorthodox” Atlantica. Due to the sparsity of poles from WAC and São Francisco – Congo during that period of Earth's history, such a revision of Atlantica still needs further investigation.

Speculation on the late Paleoproterozoic-early Mesoproterozoic supercontinent Nuna/Columbia was initiated by the recognition of widespread Orosirian orogenic belts (Hoffman, 1989), and has received much recent attention. Dozens of reconstruction models have been put forward. Here we list some representative reconstruction models, limiting our discussion to the position of WAC and proposed immediate neighbors in Nuna/Columbia. We refer readers to original publications for the details of Nuna/Columbia reconstructions. In the models of Rogers and Santosh (2002, 2009), the Atlantica supercraton idea was retained, and WAC was placed close to Amazonia and São Francisco – Congo (Fig. 13.8A). Zhao et al. (2002, 2004) compared the ages and general trends of 2.1–1.8-Ga orogens, and linked cratons in South America and Africa with Baltica to align the Paleoproterozoic basement rocks in these cratons (Fig. 13.8B). In that reconstruction, WAC faces southern Baltica so that the Archean components in Reguibat and Sarmatia can be aligned. It should be pointed out that these two models are primarily based on geological comparisons, of which the correlations would be nonunique. A more recent reconstruction of Nuna/Columbia was conducted by Zhang et al. (2012), who incorporated all available paleopoles from seven cratons and adopted the tight Laurentia-Siberia-Baltica fit proposed by Evans and Mitchell (2011) and the tight Baltica-Amazonia-WAC fit proposed by Johansson (2009) (Fig. 13.8C). The tight Baltica-Amazonia-West Africa fit, well-known as the SAMBA model, is supported by the similar 1.6–1.5-Ga rapakivi suites and anorthosite-mangerite-charnockite-granite plutons in Guiana and Fennoscandia, and also the Paleoproterozoic banded iron formations in Sarmatia, Imatoca, and Kénéma-Man domains of the three cratons (Johansson, 2009). The connection between WAC and Siberia in Zhang et al. (2012) seems to be possible based on the coeval LIP records at 1.75 Ga, 1.65 Ga, and 1.38 Ga, but the proposed connection between WAC and India currently has little supporting evidence. Pisarevsky et al. (2014), by comparing the 1.45–1.42-Ga paleopoles from Baltica and Amazonia, questioned the reliability of the SAMBA model. Instead, they argued that Amazonia and WAC were connected in Nuna/Columbia (Fig. 13.8D), but were separated from Baltica by an oceanic tract wide enough to include two opposing subduction systems (Pisarevsky et al., 2014). Bispo-Santos et al. (2008) placed North China between Baltica and Amazonia, within which Trans-North China orogen would be continuous with approximately time-equivalent orogens in Baltica (Svecofennian) and Amazonia (Ventuari-Tapajós). It is noted that despite large differences in the configurations of Nuna/Columbia in these models, WAC – Amazonia connection was commonly favored, considering the aforementioned geological and paleomagnetic evidence. However, there are models that break the WAC – Amazonia connection. For example, Chaves and de Rezende (2019) suggested a radically different model, whereby North China was located between southern WAC and northeast Amazonia, in order to match the trends of coeval 1.79–1.78-Ga mafic intrusions in three cratons, and also to align the Paleoproterozoic orogenic belts in Amazonia and North China (Fig. 13.8E). The tight Laurentia-Siberia-Baltica fit is also disrupted and WAC is assigned to occupy the core of Nuna/Columbia (Chaves and de Rezende, 2019). This model requires testing from careful, time-space comparisons of the crustal evolutions of involved cratons, and more reliable paleomagnetic data.

The existence of early Neoproterozoic supercontinent Rodinia is well-accepted, while its configuration is still highly controversial. In fact, after Li et al. (2008) published the influential and all-inclusive reconstruction of Rodinia (Fig. 13.8F), many new arguments have led to dissension rather than consensus (e.g., Santos et al., 2008; Evans, 2009; Fu et al., 2015; Ernst et al., 2016; Wen et al., 2018; Slagstad et al., 2019). Because of the inadequate paleomagnetic data and the lack of “Grenvillian” orogen in WAC, its position in Rodinia was largely constrained by the data from Amazonia, if the WAC – Amazonia connection was assumed to be valid. There remain many debates about whether the Grenville orogen in Laurentia and the Sunsás orogen in Amazonia should be continuous or facing each other, or even disconnected (Santos et al., 2008; D'Agrella-Filho et al., 2008; Evans, 2009, 2013). For instance, in Santos et al. (2008)'s model where Grenville and Sunsás orogens are continuous, southern WAC should be in contact with western Laurentia so that the mid-Paleoproterozoic Birimian and Wopmay orogens could be bridged. However, in Evans's (2009) reconstruction, Amazonia is separated from Laurentia, and the Sunsás orogen is considered to be accretionary instead of collisional. Evans (2009) also adopted the coupling of southern WAC and western Laurentia, but in that model northern WAC is attached to northeastern Amazonia (Fig. 13.8G). In this scenario, the alignment of Sassandra Belt in Man-Leo Shield and the Guri Belt in Guiana

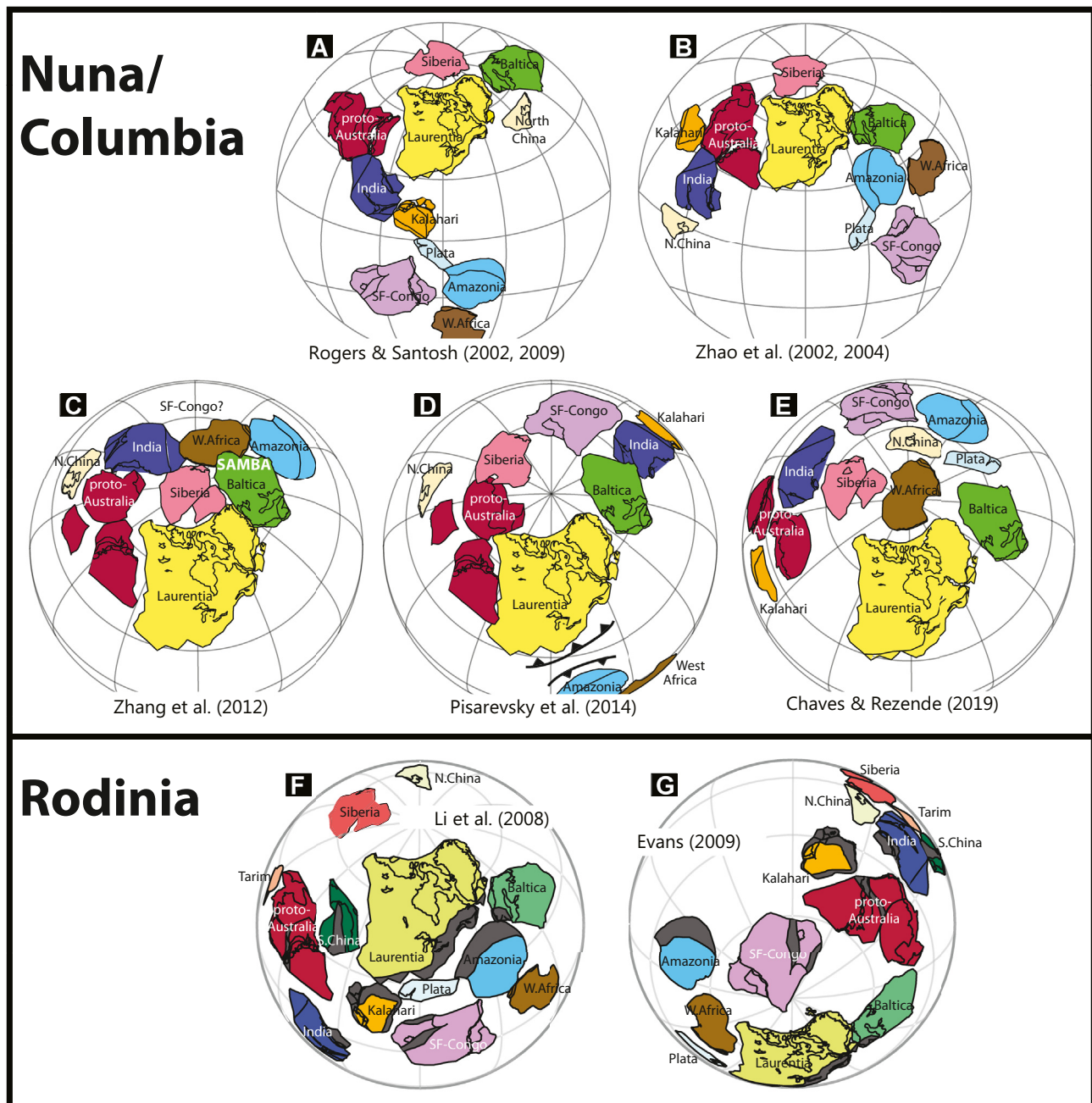


FIGURE 13.8 Various configurations of the supercontinents Nuna/Columbia and Rodinia. All projections are in present North American reference frame, with a common color scale to facilitate comparison and contrast. Gray areas indicate the 1.3–0.9 Ga “Grenvillian” orogens. (A) Rogers and Santosh (2002, 2009). (B) Zhao et al. (2002, 2004). (C) Zhang et al. (2012). (D) Pisarevsky et al. (2014). (E) Chaves and Rezende (2019). (F) Li et al. (2008). (G) Evans (2009). Modified from Evans, D.A.D., 2013. Reconstructing pre-Pangean supercontinents. *Geological Society of America Bulletin* 125(11–12), 1735–1751.

would be disrupted. Although Guri Belt could be tied in with the Sfariat Belt in Reguibat Shield, the polarity of Archean and Paleoproterozoic domains in these two shields would be twisted. The Evans (2009) model is also challenged by the need for WAC to detach from Amazonia, rotate nearly 180° and reconnect with Amazonia again in Gondwana.

After the breakup of Rodinia in the mid-Neoproterozoic, WAC experienced a series of Pan-African tectonic activities, that is, the Bou Azzer-Siroua island arc accretion to the north, collision along the Trans-Saharan Belt to the east, tectonism within the Rokelide and Bassaride Belts to the southwest, and the Mauritanide Belt to the

west (Ennih and Liégeois, 2008). WAC was welded together with other cratons to form Gondwana through these Pan-African activities. Around the Neoproterozoic-Paleozoic transition, several so-called “peri-Gondwanan” terranes (e.g., Avalonia, Cadomia, Carolina) were stripped off the margin of WAC (and/or Amazonia) and later collided with Laurentia to form Pangea (Nance et al., 2008). The opening of Atlantic Ocean in the Mesozoic separated WAC from Amazonia and brought the two cratons to their current positions.

13.7 Concluding remarks

Thus far paleogeography of WAC in the Precambrian has been poorly constrained, and the scarcity of both geological and paleomagnetic data yield notable gaps in the tectonic and kinematic history of WAC and its interactions with neighboring cratons. Since paleomagnetism provides essentially the only quantitative paleogeographic constraints, future paleomagnetic studies should take advantage of ubiquitous mafic dyke swarms recognized in WAC in order to generate high-quality and well-dated paleopoles. In addition to the efforts in new data acquisition, any future reconstruction model should proceed from a kinematic point of view, that is, hypothesizing the continuous drift history of WAC rather than proposing a single snapshot at a given time of interest. Another perspective is the significance of individual cratons in helping us understand the configuration of supercontinents, and the transitional pattern from one supercontinent to another. The ongoing discussions about how supercontinents dispersed and reassembled cyclically have inspired different geodynamic models: introversion (closure of the interior oceans), extroversion (closure of the exterior ocean), the alternation between introversion and extroversion (superocean episodes), or orthoversion (closure 90° away from the predecessor supercontinent) (Murphy and Nance, 2003; Mitchell et al., 2012; Li et al., 2019). Accurate paleogeography of poorly constrained cratons including WAC could provide a critical test for these alternative models.

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