THE CENTRAL ATLANTIC MAGMATIC PROVINCE: AN HISTORICAL PERSPECTIVE

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INTRODUCTION

The Central Atlantic Magmatic Province (CAMP) is a Large Igneous Province (LIP) of basalt that formed in the central area of the supercontinent of Pangaea, shortly before the opening of the North Atlantic Ocean in Early Jurassic time. Separation of the newly formed continents left dikes, sills, and surface flows of the CAMP along the margins and interiors of eastern North America, northeastern South America, northwestern Africa, and southwestern Europe (Fig. 1). This association of basaltic magmatism with continental rifting provides clues to features and mechanics of the mantle that might cause both events. It is not a unique association, but certainly the CAMP is one of the largest and most important LIPs to be recognized.

This overview summarizes my understanding of the CAMP and its significance, which started with regional studies from the 1970s in eastern North America and northwestern Africa, and which has continued after the recognition of the entire province. Recent efforts combine regional studies and new data toward a synthesis of the origin, ages, tectonics, and mechanisms of the CAMP, and of its ability to cause a major mass extinction.

THE CAMP NAME

A few petrologists and geochemists noted similarities between Early Mesozoic basalts on the margins of eastern North America and northwestern Africa in the 1970s, such as Weigand and Ragland (1970), May (1971), and Bertrand and Coffrant (1977). Regions within the entire basalt province had acronyms such as ENA (eastern North America), but the more comprehensive label of CAMP (Central Atlantic Magmatic Province) was first used by Marzoli and others (1999), who included dikes and sills in northeastern South America. In that year, Bill Hames, Paul Renne, and I organized a symposium about the province for the spring meeting of the American Geophysical Union in Boston (Session T02: The Earliest Magmatism of the Circum-Atlantic Large Igneous Province). At a lunch meeting afterward, the participants, who included Andrea Marzoli, discussed the issue and voted to adopt the CAMP name, and agreed to use it in future publications.

An AGU Monograph was produced by the symposium participants (Hames and others, 2003) that remains an excellent research source for information about the CAMP.

CAMP GEOGRAPHY

The map of the CAMP is defined mainly by dikes, with boundaries tentatively drawn around their farthest known extent (Fig. 1). Luckily for this work, the petrology of the CAMP dikes distinguishes them from most of the older and younger basaltic intrusions in the same regions. Swarms of related dikes tend to occur in distinct sets of dozens to hundreds that have similar orientations and field characterizations. Their genetic relationships are confirmed by petrographic (mineral) observation, bulk rock chemistry, and radiometric dates.

Sills of CAMP magma occur both in Mesozoic basin strata and also in older crustal rocks in South America and Africa. Very large tholeiite sills are mapped in Brazil and western Africa (Marzoli and others 1999; Davies and others 2017), while smaller but considerable examples are well known in the eastern USA in the Hartford, Newark, and Deep River Mesozoic basins (but not in the older basement rocks).

Mesozoic basins that preserve extrusive basalts of the CAMP total about 300,000 km^2 (McHone 2003). However, dikes and sills of CAMP that fed the basin basalts also occur across 11 million km^2 within four continents, centered upon but extending far outside of the initial Pangaean rift zone (Fig. 1). The longest dimension of CAMP is greater than 5,000 km, with several dikes greater than 500 km long, sills exceeding 100,000 km^3, and lava flows possibly larger than 50,000 km^3 (McHone 1996). If only half of the continental CAMP area was originally covered by 200 m of lava, the total volume of CAMP and ECMIP extrusive basalt exceeded 2.4 million km^3 and may be Earth's largest known subaerial flood basalt event. A very large amount may also remain in the uppermost crust in dikes and sills.

In addition, basalts of the East Coast margin igneous province (ECMIP) of North America, which cause the East Coast Magnetic Anomaly (Kelemen 1995), have a submarine area about 60,000 km^2 with perhaps 1.3 million km^3 of extrusive lavas. However, these basalts have not been yet been genetically connected to the continental CAMP and it remains possible that their formation was a different event, both in age (possibly younger) and mode of origin at the start of the new ocean crust (Benson 2003).

BASALT GROUPS

Whole-rock analyses of dikes, sills, and lavas of the CAMP tend to fall into three chemical groups (Table 1), as outlined in McHone (2000) and used by Salters and others (2003) in their discussion of the eastern North American CAMP. As shown in Table 1, the groups are characterized by average values of TiO_2: 0.62 % (low, or LTi), 1.26 % (intermediate, or ITi) and 3.21 % (high, or HTi), as well as by other components such as magnesium, nickel, and various element ratios. Some of these are used in chemical plots of the basalt types, as in Figure 2. All are tholeiites, with the LTi group mostly olivine normative, and ITi and HTi groups mostly quartz normative. As expected, phenocrysts of olivine tend to be abundant in the LTi dikes and sills, while minor interstitial quartz can be found in many of the ITi and HTi dikes, as well as early olivine in the larger intrusions.

There are also distinctions of dike swarm locations and orientations (Fig. 1). Dikes and sills of LTi basalt are nearly all found in basins and NW-trending dike swarms in the southeastern USA, whereas most of the HTi dikes are on margins of South America and Africa that were adjacent before rifting. They also tend to be in NW-SE trending dikes. LTi and HTi magmas are apparently not represented among the remnants of surface flows within the CAMP. The ITi dikes and sills are joined by large basalt flows preserved in rift basins of eastern North America and northwestern Africa. In those basin areas, the ITi dikes tend to trend NE-SW, but this group is very widespread and also has N-S dikes and other trends in other areas around the CAMP (Fig. 1).

The LTi and ITi groups can be subdivided into magma types characterized by iron, magnesium, and titanium, among other elements. The ITi divisions were recognized in an early and influential study by Weigand and Ragland (1970), who named them as HTQ (high-Ti quartz), LTQ (low-Ti quartz), and HFQ (high iron quartz) normative sub-groups. These divisions work well in eastern North America where they are abundant, but elsewhere around the CAMP area the most common type, by far, would be characterized as HTQ tholeiite. In fact, there are many slight variations within the types and groups, which reflect their primary source compositions as well as the effects of crystal fractionation, mixing, and contamination (Bertrand and Coffrant 1977; Puffer 2001; Salters and others 2003; Merle and others 2013).

CAMP GROUP AGES

In study groups at the University of North Carolina and a few other geology programs in the 1970s to 90s, there was serious speculation about the tectonic significance and relative ages of the basalt types (Ragland 1991). Several localities in the southeastern USA show ITi dikes crosscutting LTi sills and dikes (Ragland and others 1983). However, radiometric dates do not support significant age differences in eastern North America (Hames and others 2000), and more recent high-precision zircon U/Pb dates show that the Bragtown LTi sill in the Deep River Basin is 200.92 Ma, and younger than HTQ basalts that started around 201.52 Ma in eastern North America (Blackburn and others 2013). The Bragtown LTi sill age is in turn a little older than the ITi tholeiite dikes (not dated but probably LTQ types) that crosscut it in the Deep River Basin.

Moreover, the high-precision dates support the range of about 570,000 years between the earliest and latest basin basalts that was determined by Olsen et al (2003), based on basin stratigraphy correlated with Milankovitch climatic cycles. The Triassic-Jurassic boundary was found above the oldest ITi basalts in eastern North America by Cirilli and others (2009), but the end-Tr extinction horizon is still defined a meter or so beneath the oldest basin basalt (Olsen and others 2003). However, it appears that older basalts exist in Morocco (Deenen 2010) that precede the mass extinction, as well do some large sills (Davies and others 2017), and it is now generally recognized that the CAMP is the prime candidate to cause the end-Tr mass extinction (Blackburn and others 2013).

RELATIONSHIP OF THE CAMP TO PANGAEAN RIFTING

Although CAMP magmatism occurred in extremely intense but relatively brief episodes around 201 Ma, tectonic activity that led to the breakup of Pangaea was much more prolonged. The oldest rift basin sediments around the central Atlantic are early Carnian, possibly older than 230 Ma (Olsen 1997). In the southeastern USA, rifting ended before CAMP magmatism, so that sediments and basalts are spread across wide areas rather than being controlled by subsiding basins (Schlische and others 2003). It appears from seismic reflection profiles that younger Cretaceous strata are deposited directly upon the CAMP lava plains (McBride and others 1989). In the northeastern rift basins, thick Early Jurassic sediments overlie basalts (Olsen 1997), showing that rifting continued for 5 to 10 million years or more after the youngest CAMP flows, ceasing by the early Middle Jurassic (Schlische and others

2003). This diachronous rifting was once thought to correspond to the changes in dike orientations from south to north in eastern North America, but we now know the dike magmas were coeval.

The actual age of continental separation with production of the new ocean is still in question. It is generally assumed, and supported by seismic profile interpretation (Kelemen and Holbrook 1995), that the thick seaward-dipping volcanic wedge along the eastern continental margin of North America immediately preceded the formation of Atlantic Ocean crust; that is, rifting ended and drifting began. However, the oldest drift sediments in the western Atlantic margin appear to be 179 to 190 Ma (Benson 2003), or about the age of the youngest post-CAMP rift basin strata. There may be a 10-million-year gap between the CAMP magmatism and seaward-dipping wedge magmatism and new ocean crust.

MANTLE ORIGINS OF THE CAMP

Early Mesozoic dikes in eastern North America and northwestern Africa were proposed by May (1971) to radiate from a central area at the Blake Plateau, near the modern-day Bahamas east of Florida. This led to a model in which a deep-mantle plume produced not only the dikes and basalts (Morgan, 1983) but also caused the rifting of Pangaea and formation of the Atlantic Ocean crust (Storey and others 2001). Some researchers assume it, others are skeptical.

McHone (2000) pointed out that the circum-Atlantic dikes actually have trend patterns that parallel segments of adjacent rift margins of the central Atlantic (Fig. 1), and they are not radial within sets of regional dikes such as in the southeastern USA. Moreover, volcanic seamounts and islands of the Atlantic are much younger, so there is no plume track from the proposed center. As discussed above, rifting that eventually opened Pangaea started more than 30 million years before the magmatism, and the rift basins continued to develop for about another 10 million years before tectonic activity shifted to the new ocean margins (Olsen 1997). Rifting did not occur with the arrival of a plume head and the massive production of CAMP basalts, and two separate, massive events of basaltic magmatism are evidenced by the CAMP and the seaward-dipping volcanic wedge.

Weigand and Ragland (1970) described the chemical variations of the CAMP basalts as products of crystal fractionation within lithospheric magma chambers. However, it does not appear that all the chemical variations observed in the CAMP magmas can be derived through differentiation or contamination of a common mantle melt (Salters and others 2003). The upper mantle is recognized to have substantial mineralogical, chemical, and temperature variations, or heterogeneous zones from which different parent melts are derived (Shelnutt and others 2017). This model was recognized long ago by Tollo and Gottfried (1989) and does not fit a narrow mantle plume source.

Components of crustal rocks that were subducted in much older plate collision events characterize most CAMP basalts (Pegram 1990; Puffer 2001; Merle and others 2013). The research shows quite clearly that CAMP magmas are derived from different compositions of sub-lithospheric mantle, some with substantial subduction contamination, in specific regions and large geographic areas unrelated to a mantle plume center. A much better model for producing the CAMP is by the tectonic release of mantle melts, created at a critical temperature achieved from thermal insulation beneath the vast

Pangaean supercontinent; essentially the model promoted by Anderson (1994) and used by Merle and others (2013) and other researchers around the Atlantic.

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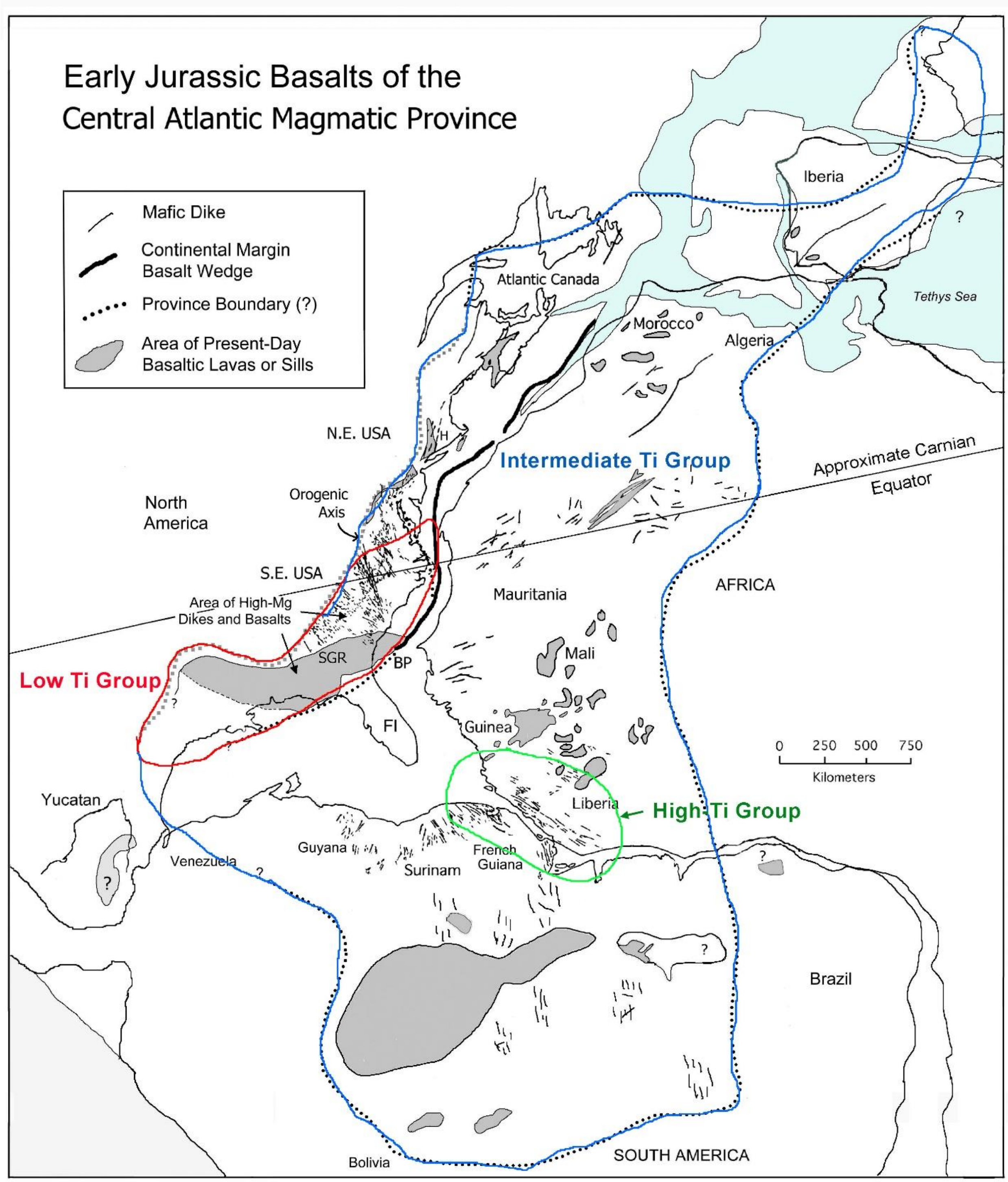

Figure 1. Generalized map for boundaries of CAMP magma types. See the text for discussion.

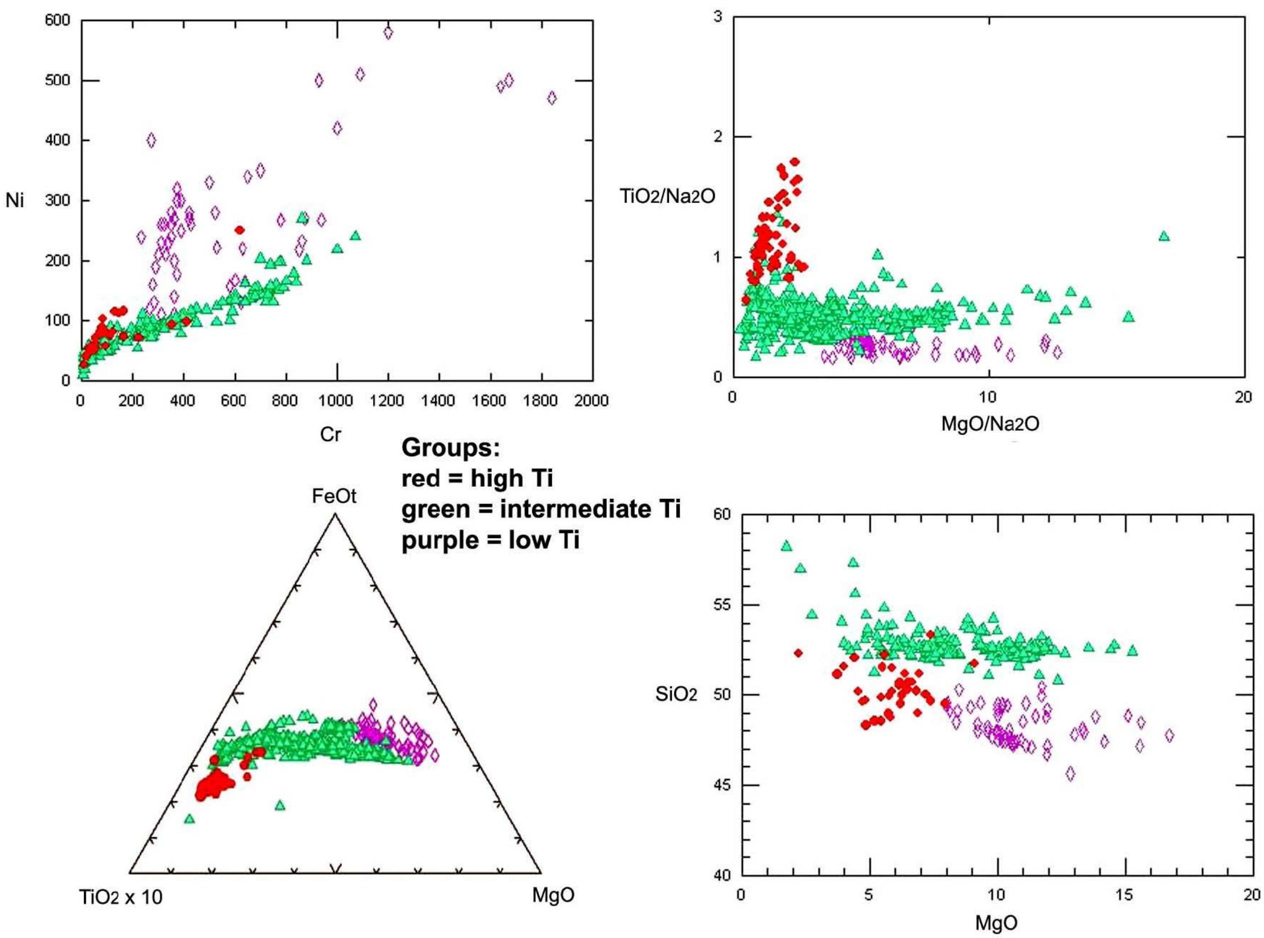

Figure 2. Chemical groups of the CAMP using components that vary with fractionation, but with group differences that indicate separate parental sources.

TABLE 1. COMPOSITIONS OF CAMP BASALT GROUPS

	LTi			ITi			HTi		
	Mean	s.d.	n	Mean	s.d.	n	Mean	s.d.	n
SiO2	48.84	2.27	142	52.61	2.54	574	51.87	2.04	60
TiO2	0.62	0.29	142	1.26	0.62	574	3.21	0.48	60
Al2O3	16.06	1.67	142	14.06	1.62	574	14.32	1.27	60
FeO*	9.92	1.08	142	10.73	1.93	627	12.14	1.65	60
MnO	0.16	0.05	142	0.18	0.03	574	0.19	0.02	60
MgO	9.46	2.44	130	6.72	3.13	574	4.11	1.16	60
CaO	10.92	1.30	142	9.92	2.11	604	7.64	1.09	60
Na2O	2.07	0.46	142	2.44	0.90	627	2.87	0.38	60
K2O	0.46	0.62	142	0.83	0.64	574	1.65	0.56	60
P2O5	0.10	0.08	142	0.17	0.11	574	0.58	0.19	60
H2O+	0.981	0.779	130	0.850	0.570	535			
CO2	0.091	0.153	133	0.124	0.671	535			
S	0.067	0.041	135	0.034	0.032	421			
F	0.023	0.055	91	0.030	0.022	411			
Cl	0.030	0.037	37	0.064	0.086	429			
Mg#	61.98	7.61		50.19	15.43		37.65		
Density	2.683	0.037		2.647	0.045		2.652		

Group names and data sources are described in the text and by McHone (2003).
LTi and ITi analyses are from Grossman et al. [1991).