

American Journal of Science

JUNE 1985

ROOTED PRECAMBRIAN RING-SHIELDS: GROWTH, ALIGNMENT, AND OSCILLATION

ALAN M. GOODWIN

Department of Geology, University of Toronto,
Toronto, Canada M5S 1A1

ABSTRACT. The distinctively large, subcircular, older Precambrian-rich Canadian, West African, and Central African shields, each with circumjacent mainly younger Precambrian-Phanerozoic rocks and broad central infilled basin, are here called ring-shields to express the resulting annular Precambrian pattern. Detailed litho-tectonic analysis by era reveals remarkably similar size, lithology, and growth patterns, the latter comprising (1) early (> 1.8 Ga) constructional stage and (2) later (< 1.8 Ga) extensional stage culminating in late Proterozoic-Phanerozoic central basin collapse. Stage-by-stage shield development is attributed to first-order thermo-tectonic events affecting a cooling, lithosphere-thickening, uniform volume Earth.

The three ring-shields are not only presently aligned on a great circle that crosses the Atlantic Ocean but have maintained this alignment during the past 600 Ma (Phanerozoic) and possibly 1200 Ma at least; alignment has been maintained despite recurrent, apparently increasing amplitude, America-Africa Wilson (open-close) Cycles each of about 400 Ma duration. A plausible explanation for sustained shield alignment during inter-shield oscillation emphasizes the role of deep, episodically-generated, sub-shield tectospheric roots to depths of 400 to 700 km or greater.

A plausible mantle-crust instability mechanism related to episodic supercontinental break-up, in turn, focuses on thermal imbalances developed in thickening crust leading to crustal reversal. Non-random cyclic plate motion, a function of terrestrial heat flow, takes the form of repeated supercontinent assembly that, however, never completely disperses, due in part to tectospheric drag caused by the deeply rooted shields. A 400 to 500 Ma-long tectonic oscillation results.

INTRODUCTION

Earth's Precambrian crust, backbone of the continents, is mainly distributed in ten continental platforms, each comprising one or more exposed shields with adjoining buried basement (fig. 1). The latter is typically disposed in broad regional undulations and dislocations including belts, arches, basins, domes, troughs, grabens, horsts, and aulacogens. Exposed Precambrian crust is concentrated in some thirty shields, each an uplifted and uncovered craton typically composed of deformed early (> 1.8 Ga) Precambrian rocks. Beyond the shield boundaries Precambrian

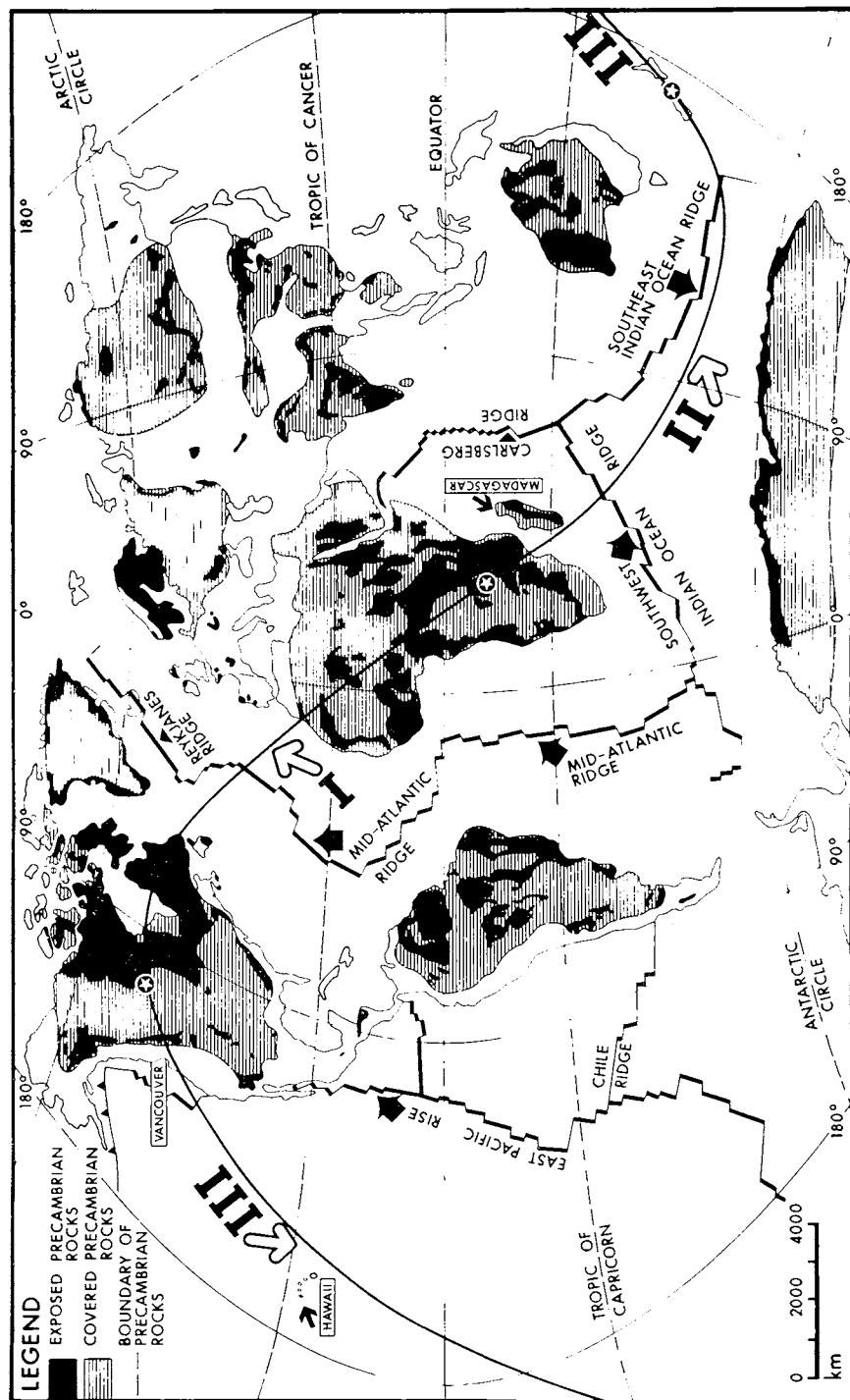


Fig. 1. General global distribution of exposed and covered Precambrian crust in present continents relative to selected ocean spreading ridges. Excluded are some small Precambrian areas as well as certain buried Precambrian crust including continental shelves. The main Precambrian continental platforms and exposed Precambrian shields are illustrated, including the generalized Canadian, West African, and Central African ring-shields with their central collapse basins of, respectively, Hudson Bay, Taoudeni Basin, and Congo Basin. The great circle along which the three ring-shields have oscillated in at least Phanerozoic time is illustrated together with equal sector component boundaries (stars).

basement typically extends laterally beneath Phaneroic cover to merge with major pericontinental fold belts (for example, Appalachian) commonly containing other smaller exposed Precambrian units. Continental platforms, in turn, represent parts of still larger, preexisting supercontinents, for example, Pangea, since rifted and drifted to present continental status.

Within this global Precambrian array three shields — Canadian, West African, and Central African — are distinctive. Each is unusually large (about 3000 km in diam), subcircular in outline, and contains a deep central basin occupied by late (< 1.1 Ga) Precambrian-Phanerozoic fill. Accordingly I propose to call them “ring-shields” to express this distinctive annular older Precambrian pattern with younger core and flanking cover. The three “ring-shields” have remarkably similar 2-stage growth patterns, an earlier (> 1.8 Ga) constructional stage and a later extensional stage culminating in central basin collapse. In common with Precambrian shields in general the “ring-shields” are inferred to have deep tectospheric roots, integral to their distinctive development.

The three “ring-shields,” which collectively subtend a third of a great circle, are not only presently aligned but have maintained this alignment throughout Phanerozoic (to 600 Ma) and probably Late Proterozoic (to 1200 Ma) time despite repeated about 400 Ma-long, mantle-induced intervening Wilson (open-close) Cycles, an oscillation also influenced by deep sub-shield tectospheric roots. This great circle also touches the Atlantic-African and central Pacific residual geoid highs, both currently centered on the equator. Precambrian shields, then, provide tantalizing insight to ancient crustal patterns somehow linked to ongoing lithosphere rifting, the basis for fruitful speculation on global shield growth, alignment, and oscillation.

By way of definition, a craton is a tectonically stable crustal block without size or age connotation. Thus the term may refer both to Precambrian platforms (continent-wide) and their contained shields (uncovered parts). Furthermore cratons of varying size, age, and duration existed during Precambrian crustal growth (for example, the comparatively small Slave Craton of late Archean age in the Canadian Shield). The following time-rock terms with boundary ages are used: Archean = > 2.5 Ga; Early Proterozoic = 2.5-1.8 Ga; Mid-Proterozoic = 1.8-1.0 Ga; and Late Proterozoic = 1.0-0.6 Ga. Equivalent terms, sparingly used in the text, are shown in figure 3. Cited ages are commonly metamorphic (rejuvenated) rather than “primary.” All calculated areas (km²) were determined by planimeter survey of referenced maps or figures.

SHIELD GROWTH

The Canadian, West African, and Central African shields, considered in turn by era (Archean, Early-Mid-Late Proterozoic, and Phanerozoic), provide comparative composition, distribution, and growth patterns. Sufficient data are provided (p. 483-508) to substantiate the interpretations (p. 508-516). Shield patterns are summarized by (1) stage, that is, early (3.8-1.8 Ga) and late (1.8-0.6 Ga) (fig. 2); (2) era (table 1); (3) shield (fig. 3);

and (4) lithologic proportions (exposed crust) by era (table 2; also entered in brackets following era designation as are key stratigraphic names used in table 1, throughout the text).

Canadian Shield

The Canadian Shield (Stockwell, 1961; Price and Douglas, 1972; Hoffman, Card, and Davidson, 1982), the major exposed part of the North American platform, is a large subcircular-orthogonal craton 3000 km in diameter and 6,864,000 km² in area (figs. 2, 3; tables 1 and 2, column 1). Archean and Early Proterozoic (Aphebian) rocks predominate. The shield is characterized by the large central Phanerozoic-filled epicratonic basin of Hudson Bay 1000 km in diameter. Three main Precambrian orogenies are recorded: Kenoran (2.6 Ga), Hudsonian (1.8 Ga), and Grenvillian (1.0 Ga); the older Mortonian (3.6 Ga) and intervening Elsonian (1.4 Ga) orogenies are expressed locally; and the Appalachian Orogeny (Taconic, Acadian, and Alleghenian Events (0.5-0.3 Ga) marks the extreme southeastern margin. The shield is bounded by Phanerozoic sedimentary on-lap except for the rifted northeast margin against Mesozoic oceanic crust underlying Baffin Bay, Davis Strait, and Labrador Sea. This well-exposed and -studied shield provides the model.

Archean Crust (25 percent) is dominated by the unusually large concave (to the northwest) Superior Craton (Goodwin and others, 1972) comprising (1) the northeastern granitoid-rich Ungava Belt with abundant high-grade (granulite) metamorphic rocks and (2) the large southwestern part characterized by east-trending, province-wide superbelt alternately rich in (A) low-intermediate grade granitoid-greenstone domains (for example, Keewatin, Timiskaming), commonly 2.7 to 2.8 Ga old but ranging about 3.0 Ga, and (B) intermediate-high grade gneissic domains with components to at least 3.1 Ga (Lac Seul) (Krogh, David, and Corfu, 1984). The high-grade granulite-rich Pikwitonei terrain (Weber and Scoates, 1978) marks the northwestern boundary (Nelson Front) of the Province. Far to the northwest (1000 km), the smaller Slave Craton (McGlynn and Henderson, 1972; Henderson, 1981) comprises granitoids associated with comparatively thin discontinuous north-trending about 2.7 Ga-old metasupracrustal belts (for example, Yellowknife). The still smaller intervening east-northeast-trending Rankin Inlet-Ennadai greenstone belt (Ridder, 1972) extends 500 km west of Hudson Bay. Other smaller Archean remnants (for example, Prince Albert Group [Schau, 1978, Frisch, 1982]) are scattered about Churchill Province. Still older Archean crust (3.8-3.5 Ga) has been dated on the Labrador coast (Nutak or Nain Province) to the east (pre-Uivak gneiss) (Collerson, Jesseau, and Bridgewater, 1975) and in the Minnesota River Valley to the southwest (Morton Gneiss) (Goldich, Hedge, and Stern, 1970), with local extensions in northern Michigan (Sims, Card, and Lumbers, 1981). The north-northeast-trending, transecting, granulite-bearing polycyclic Kapuskasing structural zone (Percival and Card, 1983), containing 1.7 and 1.1 Ga-old alkaline intrusions, provides an important structural link between Lake

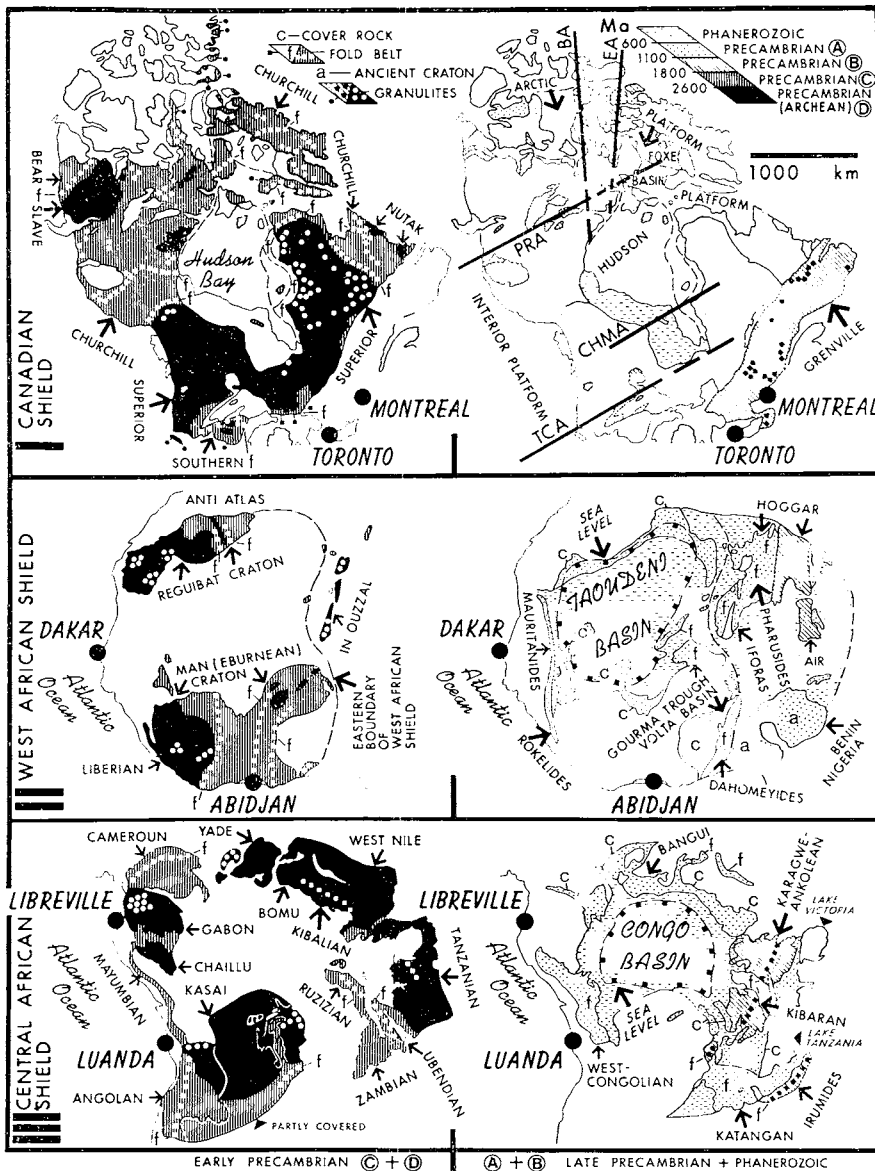


Fig. 2. Development of the Canadian, West African, and Central African ringshields by growth stages, including, on the left, early Precambrian rocks (eras C + D = 3.8-1.8 Ga) and, on the right, late Precambrian (eras A + B = 1.8-0.06 Ga) + Phanerozoic rocks. The principal Proterozoic fold belts and Precambrian granulite masses are located. The 3 central collapse basins are outlined with respect to sea level. In the Canadian Shield diagram, right side: EA = Ellesmere Arch, BA = Boothia Arch, PRA = Peace River Arch, CHMA = Cape Henrietta Marie Arch, and TCA = Transcontinental Arch. Data sources: Canadian Shield after Map 29-30, Tectonics, The National Atlas of Canada, 1973, 1:15,000,000; African shields after Tectonic Map of Africa, UNESCO, 1971, 1:15,000,000 with modifications after Metamorphic Map of Africa, UNESCO, 1978.

TABLE I

Table of development of three ring-shields by era including typical formations/depositories as cited in the text relative to the main patterns of orogeny, epeirogeny, mineral deposits, and diabase dike swarms. Selective negative epeirogenic motions refer exclusively to major central basin collapse. Fragmentation stage involved widespread extensional tectonics.

I CANADIAN SHIELD; II WEST AFRICAN SHIELD (+HOGGAR); III CENTRAL AFRICAN SHIELD

TIME SCALE MILLION YEARS	EON	ERA	FORMATIONS / DEPOSITORIES	OROGENY (SELECTED)			EPEIROGENY (SELECTED)			MINERAL DEPOSITS			DIABASE (DOLERITE) DIKES		
				I	II	III	I	II	III	I	II	III	I	II	III
0	PHANEROZOIC	CENOZOIC	▲ TA OUDENI BASIN	▲ CONGO BASIN											
22.5		MESOZOIC	▲ HUDSON BAY	▲ OBOSUM (VOLTA) ▲ SÉRIE POURPRE (H) ● BOU AZZER (OPHIOLITE) ● SÉRIE VERTÉ (H)	▲ KUNDE UNGU ▲ KATANGA ▲ SÉRIE CONGO ▲ ROYA, BUS-IMAY	APPALACHIAN PAN AFRICAN	DAMARAN								
60	PRECAMBRIAN	PALEOZOIC	● AMUNDSEN ● MINTO, BROCK	● ALEKSOD (H) ● TIDERIDJAOUNE COMP. EX (H) ● AHNET (H) ● TARKWAIAN ● GUEL EL MADID	● KIBARIDES ● IRUMIDES	GREEN VILLIAN ELSONIAN	KIBARAN	KIBARAN-BURUNDIAN							
1100		A (-ADRYNIAN)	● KEWEENAWAN ● COPPERMINE ● SEAL LAKE ● GREVILLE ● BATHURST ● BORDEN ● A. TABASCA, DUBAWNT	● ARECHCHOUM (H) ● TASSENDJANET (H) ● AFFOUT ● YETI ● BIRRIANIAN	● KIBARIDES ● IRUMIDES										
1800		B (HELIKIAN)	● ANIMIKIE (BIF) ● WOLLASTON ● FOXE ● LABRADOR (BIF) ● CORONATION ● HURONIAN ● ENNADAL ● UNION ID.	● ARECHCHOUM (H) ● TASSENDJANET (H) ● AFFOUT ● YETI ● BIRRIANIAN	● KIBARIDES ● IRUMIDES	PLATEAU SERIES									
2600	ARCHAIC	C (APHEBIAN)	● ANIMIKIE (BIF) ● WOLLASTON ● FOXE ● LABRADOR (BIF) ● CORONATION ● HURONIAN ● ENNADAL ● UNION ID.	● ARECHCHOUM (H) ● TASSENDJANET (H) ● AFFOUT ● YETI ● BIRRIANIAN	● KIBARIDES ● IRUMIDES	HUDSONIAN EBURNEAN	EBURNEAN RUZIZIAN								
3200		D EARLY	▲ TIMISKAMING (BIF) ● KEWATIN (BIF) ● YELLOWKNIFE ● PRINCE ALBERT ● UPERMIAK (L) ● IAC SEUL	▲ CHEGGA ▲ AIMSAGA (BIF) ● LIBERIAN (BIF) ● KAMBUI (BIF) ● KASILA ● IN DOZZAL (H)	● DODOMAN ● KIBALIAN (BIF) ● NYANZIAN (BIF)	KAVIRONDIAN LUIZA, ARUAN	KENORAN LIBERIAN	KASAIAN (DIBATAN)							
3600			● PRE-UIYAK (L) ● MORTON	● GANGLAN ● WATIAN ● LUANNI, LUZIAN ● BOMU	● DODOMAN ● KIBALIAN (BIF) ● NYANZIAN (BIF)	MORTONIAN OUZZALIAN	BOMU								

Legend: ● SIGNIFICANT VOLCANIC COMPONENT; ▲ ARENACEOUS; ● GREYWACKE; ■ CARBONATES; (BIF) BANDED IRON FORMATION; (H) HOGGAR; (L) LABRADOR

Mineral Deposits: Au, Cu, Co, Zn, Ni, Ti, U, Fe, Mn, Pb, Li, Sn, Ta, W, Ag, Cr, Ni, Fe, Mn, U

Diabase (Dolerite) Dikes: KARROO, FRANKLIN, MACKENZIE, MARATHON, PREISSAC (ABITIBI), MATACHEWAN

Vertical Arrows: FRAGMENTATION STAGE (upward), CONSTRUCTIONAL STAGE (downward)

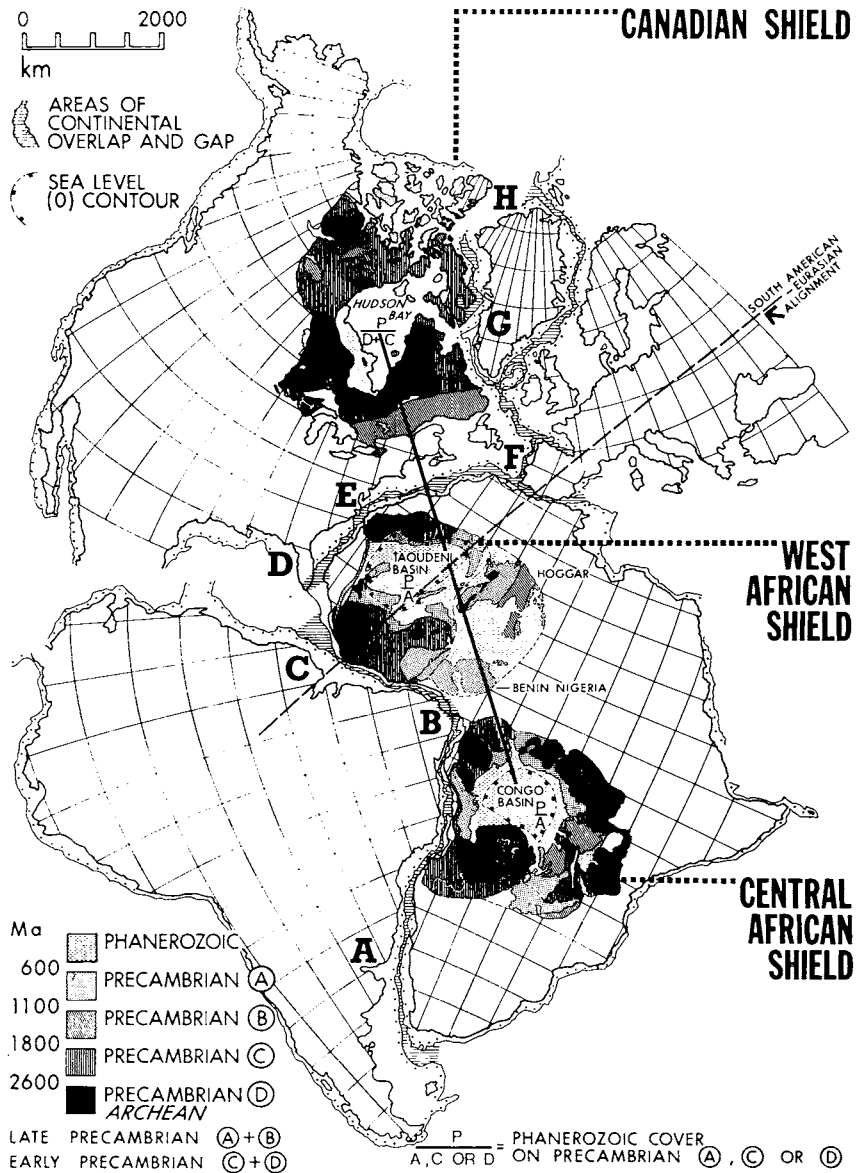


Fig. 3. Composite geology by era of the Canadian, West African (composite), and Central African ring-shields and their position on the Bullard, Everett, and Smith (1965) reconstruction of the continents about the Atlantic 200 Ma ago. Also illustrated are the North American-African triple shield alignment relative to the subordinate South American-Eurasian shield alignment.

Superior to the south and James Bay (Hudson Bay) to the north (see below).

Superior Province crust comprises (A) by lithology—granitoids: metasediments = 5:1 (Goodwin, 1978), and (B) by metamorphic facies—granulite: amphibolite: greenschist = 2:6:1 (table 2) with most of the granulite concentrated in the Ungava lobe to the northeast. Archean greenstone belts contain important Au–Ag, iron, Cu–Zn, asbestos, and nickel deposits. Archean basalt and andesite components resemble modern ocean ridge basalts and island arc andesites respectively (Capdevila and others, 1982). The belts have been variously attributed to Archean rifting (Henderson, 1981; Goodwin, 1981a) and/or subduction-related compressional regimes (Dimroth, 1982). Crustal accretion processes may have been involved. There was massive primary granitoid intrusion (McCulloch and Wasserburg, 1978) across the province during the 2.6 Ga-old Kenoran Orogeny.

Early Proterozoic (*Aphebian*) crust (34 percent) predominates in (1) Churchill Province, a vast complex region mainly underlain by gneissic

TABLE 2

Percentage composition by era of Precambrian ring-shields based on planimeter survey of source maps in figure 2. Precambrian eras as follows (International systems): D = Archean (>2.5 Ga), C = Early Proterozoic (2.5-1.8 Ga), B = Mid-Proterozoic (1.8-1.0 Ga), and A = Late Proterozoic (1.0-0.6 Ga)

	1	2	3	4	5 Average Precambrian Ring-Shield	
	Canadian	Central African	West African	Composite W. African- Touareg- Nigerian	A (cols. 1,2,3)	B (cols. 1,2,4)
Total area* (km ²)	6,864,000	5,711,000	4,103,000	6,197,000	5,559,000	6,260,000
Precambrian D (Archean)	25.4(34.0) [‡]	33.9(45.6)	14.6(26.2)	9.7(18.8)	24.6(35.3)	23.0(32.8)
Precambrian C	34.1(45.5)	19.3(26.0)	22.1(39.6)	14.6(28.4)	25.2(37.0)	22.7(33.3)
Early Precambrian (D + C)	59.5(79.5)	53.2(71.6)	36.7(65.8)	24.3(47.2)	49.8(72.3)	45.7(66.1)
Precambrian B	14.5(19.3)	7.0(9.4)	0 (0)	2.6(5.1)	7.2(9.6)	8.0(11.3)
Precambrian A	0.9(1.2)	14.2(19.0)	10.9(34.2)	24.5(47.7)	11.4(18.1)	13.2(22.6)
Late Precambrian (B + A)	15.4(20.5)	21.2(28.4)	10.9(34.2)	27.1(52.8)	18.6(27.7)	21.2(33.9)
Total Precambrian (D + C + B + A)	74.9	74.4	47.6	51.4	68.4	66.9
Phanerozoic	25.1	25.6	52.4	48.6	31.6	33.1
	100.0	100.0	100.0	100.0	100.0	100.0

* Including area of central basin in each case.

[‡] All percentages in brackets are of total Precambrian crust only (that is, excluding Phanerozoic areas).

and massive granitoids (Davidson, 1972); (2) the restricted Southern Province of the Great Lakes region (Card and others, 1972; Sims, Card, and Lumbers, 1981), and (3) the still smaller Bear Province far to the northwest (Fraser and others, 1972; Hoffman, 1980).

Older Aphebian belts and basins (2.5-2.1 Ga) are restricted to (1) the southern margin of Superior Craton (Huronian) (Card and others, 1972), (2) within Rankin Inlet-Ennadai terrane of central Churchill Province (Montgomery Lake sediments) (Bell, 1970), and (3) the southern margin of Slave Craton (Union Island) (Fraser and others, 1972). They are characterized by thick sandstone-pelite-conglomerate sequences with minor flood basalts, typically disposed in off-craton thickening wedges. Placer uranium deposits are important in the Huronian Supergroup.

Younger Aphebian mobile belts and basins (2.1-1.8 Ga) are widespread (fig. 2A; see also Goodwin, 1974, fig. 5 for individual belt identification and description). The belts lie in three groups arranged at increasing distance from Superior Craton: (1) Circum-Superior group comprising Animikie-Marquette and Mistassini-Otish basins, Labrador Trough, Cape Smith, and Belcher belts, Sutton Inlier, and Fox River and Thompson belts; (2) Trans-Churchill group which arcs across Churchill Province concave to the south, including Wollaston, La Ronge-Lynn Lake, Flin Flon, Rankin Inlet-Ennadai, Dorset, Foxe, and Committee belts; and (3) Wopmay group, mainly to the west of Slave Craton. Typical Early Proterozoic belts comprise thin shelf facies (orthoquartzite-conglomerate-carbonate-banded iron-formation) and thicker off-shelf slope-rise turbidite facies (graywacke-mafic-ultramafic volcanics and intrusions). Major iron ore deposits lie in Animikie-Marquette and Labrador-Belcher sequences. Significant Cu-Zn-Au deposits occur in some greenstone belts, for example, Flin Flon. Major Ni-Cu-Pt-Cd ores occur in the Sudbury Lopolith (Card and others, 1972; Krogh, McNutt, and Davis, 1982) of Southern Province, and major Ni-Cu ores occur in the Thompson and Cape Smith belts.

The following typify current Early Proterozoic tectonic interpretations, most of "early extension-later compression"-type.

1. *Circum-Superior belts*.—A. Apehbian successions of the Lake Superior-Lake Huron region have been varyingly ascribed to (1) intracratonic deposition in rifts controlled by a fundamental basement boundary between late Archean (2.6-2.8 Ga) granitoid-greenstone terrane to the north and early Archean (3.5-3.0 Ga) gneissic terrane to the south, followed by terminal compression involving northward transport of the gneissic terrane and its deformed cover (Sims, Card, and Lumbers, 1981), (2) deposition on a passive continental margin followed by north-dipping subduction of oceanic-crust (van Schmus, 1976), and (3) deposition on a rifted continental margin followed by continent-continent collision during a south-dipping subduction zone (Cambray, 1978).

B. The Labrador Trough is a south-trending splay off the main Churchill Province. Trough development (Wardle and Bailey, 1981) is attributed to an ancient continental shelf-rise-slope system developed on

the western passive margin of a proto-oceanic rift system. Earlier shelf development to the west was succeeded eastward by later voluminous mafic volcanics and thick pelites, these facies now in thrust fault contact. Models for plate closure and collision have been proposed (Wilson, 1968b; Gibb and Walcott, 1971; Thomas and Kearey, 1980). On the other hand, Dimroth (1970, 1981) favors ensialic development in the Trough.

C. A similar 3-stage rifting model has been proposed for the coeval Belcher Group (Ricketts and Donaldson, 1981) at the western margin of Superior Craton in Hudson Bay: (1) growth on an initial rift-induced shallow marine platform, (2) a second miogeoclinal stage composed of carbonate platform-to-basin sequence, prograding carbonate buildup with associated banded iron-formation (Kipalu Fm), and extensive submarine volcanism to form a thick volcanic arc, and (3) a rift basin, including the Richmond Gulf Graben or aulacogen (Chandler, 1982), filled by submarine fan facies turbidites and later arkosic red beds. Subsequent isoclinal folding during the terminal Hudsonian Orogeny accompanied presumed ocean-closing and continental collision tectonics.

D. Baragar and Scoates (1981), modifying Wilson (1968b), Gibb and Walcott (1971), and Burke and Dewey (1973), proposed a plate-tectonic origin for the entire circum-Superior belt system. Stretching of a sialic crust produced fracturing around the preexisting stable craton, the ancestral Superior Province, followed by necking along fractures to produce annular sedimentary troughs. With continued separation, volcanism was initiated to onlap onto the earlier sediments. Subsequently the southern part (Lake Superior-Grenville) of the encircling rift system expanded at the expense of the northern part of an opening ocean with subduction at the continental edge. The northern rift system (Labrador Trough, Cape Smith, Belchers, Foxe River, Thomson) ceased activity and eventually closed with terminal deformation.

2. *Trans-Churchill belts*.—Broad (+ 300 km) extensive ensialic Hudsonian mobile belts incorporating both remobilized Archean basement and younger sedimentary cover are typified by the northeast-trending Cree Lake Zone in the Lake Athabasca region, Saskatchewan (Lewry and Sibbald, 1980; Lewry, Stauffer, and Fumerton, 1981). This zone comprises a core of remobilized Archean basement with narrow flanking belts (for example, Wollaston) of deformed Early Proterozoic shelf-type supracrustal rocks. Granitoid intrusions are common in both the core and the flanking belts. Within the Cree Lake Zone, Archean basement and Early Proterozoic cover developed complex mobile gneiss-migmatite infrastructures that straddle the amphibolite-granulite facies boundary. Heat transfer to higher crustal levels was varyingly achieved by (1) voluminous vertical and lateral flowage involving widespread development of nappes and gneiss domes with recumbent migmatite lobes and other flat-lying structures, and (2) vertical flow (diapirs) into upright mantled gneiss domes. Compression was important at least in later tectonic stages. The Cree Lake Zone changes eastward (Lewry and Sibbald, 1980) to low-medium grade granitoid-greenstone domains (La Ronge-Lynn Lake and Flin Flon

belts). This could imply either subduction tectonics with involvement of substantial oceanic crust, or, alternatively, in the absence of recognized sutures, limited rifting with later eastward Superior Province-directed compression (Lewry and Sibbald, 1980). Many other medium- high-grade mobile belts of Churchill Province are considered to reflect similar zones of crustal mobility. This would imply widespread upward-outward (vertical-lateral) tectonism in development of Churchill crust (see below).

3. *Wopmay Orogen*.—The 1.9 to 1.8 Ga Wopmay Orogen (Bear Province) is interpreted by Hoffman (1980) as representing a complete Wilson (open-close) Cycle. At about 1.9 Ga, a system of rifts led to continental break-up along the western margin of Slave Province with deposition of a west-facing (present coordinates) continental terrace-and-rise system. It is proposed that alternately west- and east-dipping subductions of oceanic lithosphere led to (1) collisional accretion of an arc-bearing microcontinent, followed by (2) major extrusive-intrusive activities to the west (Great Bear Batholith). A terminal collision to the west resulted in a wide swath of conjugate (northeast-northwest) transcurrent faults. Consequently Slave Province was transported eastward against northwestern Churchill Province, with resulting indentation and faulting. In the absence of recognized relict oceanic crust the subduction process remains unproven, however, plausible. Despite this uncertainty, the eastward tectonic transport of Wopmay Orogen is well-expressed in the Cloos Nappe (Hoffman and Pelletier, 1982), a recumbent, eastward verging, refolded anticlinorium near the eastern side of the Hepburn metamorphic-plutonic belt, and other stacks of thrust nappes showing west-to-east overlap (St. Onge, King, and Lalonde, 1982). The Kilohigok Basin to the east (not illustrated), which developed as a splay of the Taktu Aulacogen of Wopmay Orogen, is the depositional site of the Goulburn Group (Campbell and Cecil, 1981), an intracratonic basin intimately related to the development of Wopmay Orogen.

Fold belts of Churchill Province are characterized by closely spaced parallel folds commonly steep and isoclinal. Typical tectonic transports have been cratonward, that is, toward Superior Province. For example, the pattern of folding in the Circum-Ungava fold-belt is one of increasing deformation away from the Superior Craton (Davidson, 1972). A key segment, the Labrador Trough, with a present width of 130 km, has undergone an estimated minimum 50 percent shortening (Dimroth, 1981); in consequence, catazonal hinterland granulite-rich allochthons are now exposed alongside mesozonal-epizonal hinterland and foreland rocks. Other granulite-rich domains in the province (fig. 2, table 3), especially large tracts on the northern perimeter (Baffin Island, Ellesmere Island, Boothia Peninsula, Queen Maud Gulf, Great Slave Lake), probably reflect similar vertical-cratonward that is, southeastward, transport of deeper crust. In Southern Province, large, northward-directed recumbent folds (nappes) have been identified in east-central Minnesota (Holst, 1984). The net amount of horizontal transport, crustal interleaving, and stacking attendant on repeated infrastructural mobilization of Archean basement-

Early Proterozoic cover across Churchill Province constitutes an impressive component of shield growth and cratonization.

That most granitoid intrusions in Churchill Province represent rejuvenated Archean crust, earlier interpreted on field relations (Fraser, 1978; Lewry and Sibbald, 1980; Jackson and Morgan, 1978), finds support in preliminary Sm-Nd isotopic systematics (McCulloch and Wasserburg, 1978).

Mid-Proterozoic (Helikian) crust (15 percent) includes both (1) fault bounded (rifted) basins and (2) a single major mobile belt (Grenville).

1. Fault-bounded basins characterized by redbed-quartzite-bimodal volcanic assemblages include both older Helikian (1.7-1.4 Ga) epicratonic basins (A) Athabasca, and (B) Dubawnt, and younger Helikian pericratonic basins, (C) Coppermine River, (D) Bathurst Inlet, (E) Borden, and (F) Lake Superior.

A. Epicratonic basins of the Athabasca region (Raemaker, 1981) formed as a series of pull-apart structures in an episode of transcurrent faulting during Hudsonian events, possibly collisional, in eastern Saskatchewan. Earliest Helikian (1.7-1.6 Ga) basins (for example, Martin) represent small wrench-fault units formed along strongly active faults with unmetamorphosed redbed infill. The younger (1.5 Ga) Athabasca basin resulted from less active rejuvenation of the same northeast-northwest-trending fault systems. This basin was infilled by a molasse wedge (Athabasca Group), fluvial in the east but lacustrine-marine in the west, shed from a quartz-rich metamorphic source to the east, possibly related to Hudsonian collision (Raemaker, 1981). The basin fill mainly represents first-cycle orthoquartzites. It comprises regolith and basal conglomerate, four marine transgressions, and a thick fluvial regressive wedge, all products of rapid accumulation in a tectonically active area. Large U-Ni ore deposits are present at and near the base of the Athabasca basin.

B. In the Baker Lake region, 600 km to the north-northeast, the 1.7 Ga-old Dubawnt Group (Fraser and others, 1970; Blake, 1980; Heywood and Schau, 1981), a similar quartzite-rich basin fill comprises a lower

TABLE 3

Proportions (percent) of metamorphic facies in selected structural provinces of and complete Canadian Shield (based on planimeter survey by A. M. Goodwin of Map 1475A, Canada Geol. Survey)

Province (Major Orogeny)	Superior (Kenoran, 2.6 Ga)	Churchill (Hudsonian, 1.8 Ga)	Grenville (Grenvillian, 1.0 Ga)	Total ¹ Canadian Shield
Area (Km ²)	1,572,000	2,128,000	753,000	5,560,000
Granulite	22.4	27.7	25.1	23.7
Amphibolite	66.2	55.2	66.0	59.7
Greenschist	11.0	11.8	8.9	11.9
Subgreenschist	0.4	0.9	0	1.8
Unmetamorphosed	0	4.4	0	2.9
	100.0	100.0	100.0	100.0

¹ All 7 structural provinces.

sequence of fault-bounded redbeds (South Channel and Kazan Fms), a middle sequence of alkalic and felsic volcanic rocks (Christopher Id Fm), and an extensive upper sequence of quartz sandstone, conglomerate, and dolomite (Thelon Fm).

Smaller late Helikian (about 1.2 Ga) pericratonic redbed-quartzite-volcanic-filled basins are present at (C) Coppermine River (Kerans and others, 1981) and (D) Bathurst Inlet (Fraser and others, 1970), both in the Wopmay Orogen region. Both accumulated in basins associated with northeast- and northwest-trending conjugate wrench faults that formed, according to Hoffman (1980) and Hoffman and St. Onge (1981), as a result of Aphebian plate collision. (E) The coeval pericratonic Borden Basin, 1200 km to the northeast, was initiated as an aulacogen, about 1250 to 1200 Ma ago, in the North Baffin Rift Zone, a 1200 km-long rift along the northern edge of the Canadian-Greenland Shield associated with the opening of Poseidon Ocean, the Proto-Arctic Ocean (Jackson and Iannelli, 1981). Bylot Supergroup, up to 6 km thick, includes earlier narrow fault-controlled channel and broader alluvial braidplain deposits, an overlying sandstone-shale delta fan complex and subtidal shales (Eqalulik Gp), and succeeding stromatolitic shelf carbonates (Uluksun Gp) accumulated during subordinate faulting. Major faulting accompanied accumulation of still younger sandstone, shale, carbonates, and conglomerates in alluvial fans to subtidal environments (Nunatsiag Gp). The uppermost intertidal sandstones and subtidal shelf quartz arenites are disconformably overlain by Paleozoic strata.

Significantly, post-Aphebian (1.8 Ga) to recent northwest-trending faults abound in Baffin Island resulting in the formation of numerous horsts and grabens (Jackson and Iannelli, 1981). This type of activity is considered to be a key structural element in the later development of Hudson Bay (see below).

Finally in the Lake Superior region, 2500 km to the south, (F) Keweenawan rocks accumulated within a + 2000 km-long Mid-Continent Rift system (Wallace, 1981; Klasner, Cannon, and Van Schmus, 1982). Deposition of Sibley quartz arenites began about 1537 Ma ago (David and Sutcliffe, 1984). Between 1140 and 1120 Ma active rifting, involving 60 to 90 km of crustal separation, initiated rapid extrusion of enormous volumes (+ 400,000 km³) of subaerial tholeiitic basalts with sparse felsic associates in the Lake Superior region. Interflow lithic sandstones and polymictic conglomerates are subordinate. Keweenawan intrusions, mainly gabbroic, are dominated by the Duluth Complex. Local alkaline intrusions of ages 1.7 and 1.1 Ga are present. Later moderate compression steepened the sides of the basin to form the modern Lake Superior Syncline. Major Cu(-Ag) ore deposits lie in Keweenawan rocks. Similar coeval rift-associated flood basalt assemblages are present at Seal Lake in southern Labrador and in Coppermine Homocline, Bear Province.

2. The Grenville Province, 300 to 600 km wide and 2000 km in exposed length, is an eroded orogenic belt distinctive for its widespread high-grade metamorphism (table 3), complex deep-level structures, and

abundant anorthosites (Wynne-Edwards, 1971, 1976; Baer, 1976, 1981). The belt evolved over a period of about 600 Ma (1700-1050 Ma) and is especially marked by the culminating Grenvillian Orogeny (1100-1000 Ma). Metamorphism and deformation are considered to be predominantly ensialic involving both older continental crust and its platform cover (Wynne-Edwards, 1976). Although most (about 60 percent) of Grenville province is underlain by quartzofeldspathic gray gneiss representing rejuvenated basement, autochthonous metasedimentary assemblages, 3 to 9 km thick, of shallow marine or continental affinity (marbles, calc-silicates, metapelites) are present (for example, Grenville Supergroup). Vast sheet-like anorthositic intrusions are distributed in north-trending alignments in which the intrusions become progressively younger to the south, all within the 1480 to 1050 Ma interval. Gibb and Walcott (1971) proposed a continental rifting-collision model, and Wynne-Edwards (1976) a ductile spreading ensialic model. The prevailing deformation across the entire province is characterized by a pattern of northwest-verging and northeast-trending folds (Wynne-Edwards, 1976). This major overall northwest tectonic transport has resulted in widespread exposure of presumed depth-dependent catazonal granulite and anorthosite. The prevailing level of metamorphism in Grenville crust indicates an average depth of erosion of 15 km across the province (Wynne-Edwards, 1972). Many juvenile Grenvillian age granitoids were added to the crust (McCulloch and Wasserburg, 1978).

Prevailing Grenvillian northwest tectonic transport pattern is expressed in the Parry Sound region, Ontario, where Davidson, Culshaw, and Nadeau (1982) document the presence of numerous, continuous tectonic zones (thrust belts) along which predominantly northwest-directed movement has taken place. The zones outline distinctive domains, products of tectonic emplacement of allochthonous crustal segments by northwest-directed shear and thrusting. Such a mechanism is compatible with the Tibetan plateau environment of the Himalayan-type model for development of thickened crust (Burke and Dewey, 1973).

The recognition of large-scale nappe structures and thrust faults in Grenville Province is further supported by detailed geochemical studies of the Tallan Lake Sill, a 1246 ± 24 Ma-old structurally overturned amphibolite-syenite intrusion in northern Grenville Province (Heaman, Shaw, and McNutt, 1982).

Late Proterozoic (Hadrynian) pericratonic basins (1 percent) are mainly restricted to the northern shield. In the Coppermine River area of the western Arctic, sandstone, shale, argillite, stromatolitic carbonate, and gypsum of the Rae Group unconformably overlie 1.1 Ga-old Coppermine River Group supracrustals (Fraser and others, 1970). Rae Group strata are tentatively correlated with sediments in the Brock and Minto Inliers to the north. The thickness of Hadrynian sediments in the encompassing Amundsen Basin is 1300 to 3500 m. The 675 Ma-old Natkusiak flood basalts, up to 350 m thick, complete Hadrynian accumulation in this area.

To the northeast, the pericratonic Franklinian Fold Belt contains more than 12 km thick of latest Proterozoic-lower Paleozoic sediments (Kerr, 1980). Continental margin sediments were disposed about a fundamental east- to northeast-trending hinge line in the Melville Sound-Jones Sound region separating thin platform sediments overlying Early Proterozoic-Archean crystalline basement to the south from thick slope-rise sediments to the north. The basin has been deformed into several major fold belts dominated by decollement structures formed by south-to-southeastward overriding of the older Ellesmerian Orogen. Similar Mid-Late Proterozoic off-shield assemblages are present in the pericontinental Cordilleran and Appalachian Fold Belts.

Phanerozoic Crust (25 percent). In the Arctic the younger Sverdrup Basin of upper Paleozoic-Mesozoic age developed upon the folded remnants of the Franklinian Fold Belt. The Sverdrup Basin, created by active faulting, became a broad sag controlled by reactivated deep-seated predominantly east-northeast trending extension faults controlled by older structures (Kerr, 1980).

Hudson Bay with both Foxe Basin in the north and Moose River Basin in the south contains a maximum thickness of 2000 m (Douglas, 1970) of Paleozoic (Ordovician-Silurian-Devonian) and local thin Cretaceous strata. These unconformably overlie Archean (beneath Moose River Basin to the south) and Early Proterozoic (beneath Hudson and Foxe basins to the north) basement (see Douglas, 1970, fig. IV-37, p. 138). Thus the erosional interval represented by the unconformity at the base of Hudson Bay represents a remarkable 1350 Ma hiatus (1800-450 Ma). The origin of Hudson Bay is ascribed to block faulting and crustal sag. It is a product of recurrent conjugate (northwest-northeast) extensional fault systems to which the older Borden Basin (about 1200 Ma), the coeval Franklinian and Foxe basins (600-300 Ma), and the younger Sverdrup Basin (300-100 Ma) are related. The north-trending older Precambrian Boothia and Ellesmere Island arches (fig. 2) may have formed a natural boundary to further collapse of Hudson Bay on the west, as did the Ungava lobe of Superior Province on the east, the northeast-trending major Paleozoic continental uplifts — Peace River Arch (also separating Athabasca and Thelon basins) on the north, and the Transcontinental Arch on the south (fig. 2A; Bally, 1980, fig. 9A). The northeast-trending Cape Henrietta Maria Arch (fig. 2A) similarly separates Hudson Bay and Moose River Basin. All these basins, together with the 1.2 Ga-old Mid-Continent Rift and Kapuskasing Fault systems to the south (Percival and Coe, 1981, 1983), may represent successive episodes in continuing continental extension and fragmentation which eventually brought about the evolution of the present Arctic (Cretaceous) and Atlantic (Jurassic) ocean basins as well as branches within the continent.

Continental mafic dikes, products of shield-wide tension (Fahrig, Christie, and Jones, 1981; Fahrig *in* Goodwin and others, 1972), developed at 1200 Ma (Mackenzie, Logan) and 675 Ma (Franklin). Older swarms — 1900 to 2000 Ma (Marathon), 2150 Ma (Abitibi, Preissac), and 2630 Ma

(Matachewan) — are concentrated in the southern and eastern shield. Four such swarms (Payne River, Mistassini, Marathon, and Molson) may be related to early Proterozoic Circum-Superior spreading episodes, the dikes representing failed arms of triple junctions (Fahrig and others, 1984).

Metamorphism.—The Canadian Shield contains a spectrum of metamorphic facies from subgreenschist to granulite (Fraser and Heywood, 1978). Granulite facies is present in all structural provinces and collectively represents almost 24 percent of exposed Precambrian crust (table 3). Major granulite belts and masses are concentrated in Superior (22 percent), Churchill (28 percent), and Grenville (25 percent) provinces, products respectively of Kenoran (2.5 Ga), Hudsonian (1.8 Ga), and Grenvillian (1.1 Ga) orogenies. The presumed depth dependent granulites mainly represent catazonal crust tectonically transported upward to present exposure alongside mesozonal-epizonal crust. The total implied shield-wide tectonic transport is immense. The dominance (about 60 percent) of amphibolite facies (mainly upper amphibolite) corresponds to a net shield-wide epeirogenic uplift and erosion of 10 to 15 km. This uplift may be reasonably ascribed to long-term buoyancy of increasingly granitoid-rich crust during shield growth. Lower grade facies terranes, notably in southern Superior Province and in Slave Province, represent those parts of the shield that have undergone minimal net uplift and erosion.

West African Shield

The West African Shield (Bessoles, 1977; Bessoles and Trompette, 1980; Choubert and Faure-Muret, 1971; Saggerson, 1978; Black, 1980; Affaton, Sougy, and Trompette, 1980; Clauer and others, 1982; Cahen and others, 1984) is a large ovoid craton 2300 km (north-south) by 1700 km in dimension and 4,103,000 km² in area (figs. 2, 3; tables 1 and 2, column 3). Late Proterozoic (19 percent) and Phanerozoic (44 percent) rocks predominate, these marking the presence of the large (1000-1500 km diam) central Late Proterozoic-Phanerozoic-filled Taoudeni Basin and smaller Volta Basins to the southeast. Adjoining to the east are the Touareg (Hoggar) Shield (the Iforas-Pharuside-Hoggar-Air collective of fig. 2B), an amalgam of narrow, north-trending shear bounded domains or microplates, and, to the south, the Benin-Nigeria Shield (fig. 2). The resulting composite (West African-Touareg-Benin Nigeria) is 6,197,000 km² in area (table 2, column 4). Three main orogenies are recorded: Liberian (2.6 Ga), Eburnean (1.9-1.6 Ga), and Pan-African (0.6-5 Ga); earlier Ouzalialian (about 3.3 Ga) and possibly the intervening Kibaran (1.1 Ga) orogenies are recorded in the Touareg Shield. Both the West African Shield proper and the designated Precambrian composite are bordered by Pan-African (Pharusian, Dahomeyan, Rockelide, Ougartan) and Hercynian or Variscan (Mauritanides, Rockelides, Anti-Atlas) fold belts except where locally rifted against Mesozoic oceanic crust of the Guinean Gulf to the south.

Archean rocks (15 percent) of the West African Shield are concentrated in two older Precambrian (that is Archean-Early Proterozoic)

cratons (dorsales), Reguibat to the north, and Man (Eburnean or Leo) to the south (Bessoles, 1977) (fig. 2). In western Reguibat (Tiris), the Amsaga Group includes the older Rag el Abiod migmatite complex (to 3.02 Ga) and the younger Saouda metamorphic assemblage composed of quartzofeldspathic rocks, Al-schists, amphibolite, marble, and itabirite or banded iron-formation (BIF). To the east, Ghallam gneiss have yielded an unconfirmed date of 3.27 Ga (Cahen and others, 1984). The Fderik (Fort Goureaux) iron deposit contains BIF in an epizonal allochthon of possible Early Proterozoic age (Bessoles, 1977). The Chegga Group of central Reguibat includes abundant quartzofeldspathic gneiss, amphibolite, marble, and quartzite but lacks BIF. An uncertain age of 2.71 Ga has been obtained on a syenite intrusion (Cahen and others, 1984). Typical metamorphism across Reguibat Craton decreases from granulite facies in the west (Tiris) to amphibolite facies in the east. A major orogenic event (Liberian) occurred about 2.7 Ga ago (see below).

To the south, western Man (Liberian) Craton is characterized by migmatite, charnockite, dioritic gneiss, pyroxenite, amphibolite, and BIF, products of the Liberian cycle of events (3.0-2.6 Ga). The type Liberian assemblage (Bessoles, 1977) features medium-high grade crystalline basement providing dates of 3.0 Ga (Cahen and others, 1984) and meta-sedimentary cover, mainly pelitic schist, calc-schist, amphibolite, and BIF with recorded older dates of 2.7 Ga. Two greenstone sequences are recognized, the younger characterized by extensive BIF. In Sierra Leone to the west (Williams, 1978a, b) the Archean is divided into the north-trending BIF-bearing Kambui schist (greenstone) belts of about 2.7 Ga, themselves resting on basement dated at 3.0 Ga (Cahen and others, 1984), the Kasila Group Mobile Belt, a northwest-trending medium-high grade supracrustal sequence at the southwest margin of the Craton, which reportedly underwent high-grade metamorphism at 2.7 Ga, and the associated lower grade klippen, the Marampa Group of possible pre-Liberian age. The Kambui-Kasila boundary is a zone of high strain and thrusting. The last major event (Liberian orogeny) is 2.8 to 2.6 Ga (Hurley, Fairbairn, and Gaudette, 1976; Clauer and others, 1982). Hedge, Marvin, and Naeser (1975) concluded that sialic crust formed about 3.2 Ga ago and that the Liberian supracrustal rocks accumulated sometime between 3.2 and 2.7 Ga. The entire assemblage was then affected at about 2.7 Ga during the Liberian event. The Nimba BIF-bearing formation, which closely resembles the Kambui schists, contains major iron and manganese deposits. Significant Cr and Au deposits are present in nearby greenstone belts.

Finally, in the adjoining Touareg Shield to the northeast, the In Ouzzal-Iforas block of dominant east-northeast structural trend includes Ouzzalian gneiss, granulite, and charnockite dated at 3.2 and 3.0 Ga (Bertrand and Lasserre, 1967; Saggerson, 1978), the latter possibly representing the age of granulite facies metamorphism. Granulites of the Red series in the central Hoggar provide an age of 3.5 Ga (Cahen and others, 1984).

Early Proterozoic (22 percent) rocks represent both supracrustal sequences and rejuvenated pre-Eburnean (2150 ± 100 Ma) basement (Bessoles, 1977; Black, 1980). In eastern Reguibat to the north, Yetti volcano-detritic and volcanic rocks of common greenschist facies are intruded by high-level Eburnean granitoids with closely associated ignimbrites (Aftout). Sheets of unmetamorphosed Lower Proterozoic sandstone-conglomerate have been tectonically thrust southwestward over the Yetti Group. In central-eastern Man (Eburnean), to the south, broadly distributed north-northeast-trending early Proterozoic belts form the well-known 2.3 to 2.1 Ga-old Birrimian greenstone assemblage (Black, 1980; Cahen and others, 1984); the belts comprise a lower volcanic-flysch sequence distributed in deep, elongated troughs (type I) with Fe, Mn, Au, and, to the east, local diamond deposits, and an upper arenaceous-rudaceous sequence distributed in broad, shallow basins (type II). Type I Birrimian belts are typical greenstone assemblages with metabasalts, felsic tuffs, quartzite, and phyllite including persistent gondite (Mn) horizon, commonly overlain by flysch. Type II Birrimian felsic volcano-sedimentary deposits feature graywacke, quartzite, and chlorite schists (Black, 1980). To the east (Ghana) volcanic rocks are more abundant, especially in the upper Birrimian. Two isolated inliers to the northwest, Kenieba and Kayes, comprise Birrimian volcanodetritic sequences with flysch intruded by Eburnean granitoids (Black, 1980). Eburnean orogeny, widespread in northern Africa, is well-dated in both Reguibat and Man cratons by large granitoid intrusions about 2.0 Ga old (Clauer and others, 1982). The time span of the Eburnean is set at 2.1-1.9 Ga (Cahen and others, 1984). Younger ages, up to 1.7 Ga, represent younger events as well as possibly cooling effects. Large north- to northeast-trending transcurrent faults with blastomylonites developed late in the Eburnean orogeny. Local kyanite records high metamorphic pressures.

In both Reguibat and Man cratons, unconformably overlying coarse-grained, undeformed to moderately deformed, non- to slightly metamorphosed post-orogenic molasse (Guelb el Hadid and Tarkwaian, respectively) are broadly distributed in up to 4 km-thick units. Locally, this cover is cut by Eburnean granitoids (1.9-2.1 Ga).

In the Touareg Shield to the northeast the large Tassendjanet nappe structure (Bertrand and Caby, 1978) comprises mainly quartzite, Al-schist, marble, gneiss, and amphibolite of the Tassendjanet Formation, intruded by Eburnean granitoids. In the Aleksod area (Bertrand and Lasserre, 1976) of east-central Hoggar, the "Série a L'Archchoum" comprises pelitic gneiss, marble, and quartzite varyingly dated at 2222 to 1940 Ma. Practically throughout the Hoggar a Tassendjanet-Arechchoum-type basement can be defined, leading Bertrand and Lasserre (1976) to conclude that the whole of the Hoggar is underlain by Eburnean (about 1.9 Ga) or older basement.

Mid-Proterozoic rocks are virtually absent in the West African Shield proper. This 700 to 900 Ma-long gap, eloquently labelled "La Grande Lacune", covers the interval from the 1.9 to 1.7 Ga-old Eburnean base-

ment to the 1035 Ma-old lowermost basement cover (Série a Stromatolites) (Clauer, 1976). The only possible sedimentary representative in this interval is the Tarkwaian-Guelb el Hadid molasse, of which the Tarkwaian may, in fact, be 2180 to 2000 Ma old (Bonhomme and Bertrand-Safarti, 1982).

Mid-Proterozoic rocks are, however, well-developed in the Pharusian Belt (Caby, Bertrand, and Black, 1981; Bertrand and Caby, 1978) lying between the eastern edge of the West African Shield and the polycyclic central Hoggar (fig. 2) and dominated by rocks formed between 1750 and 600 Ma. The Ahnet Formation comprises a thick (> 3 km) accumulation of fine grained orthoquartzites with rare pelitic intercalations and, on top, stromatolitic dolomite together with metamorphosed equivalents to the west. This preceded emplacement of the extensive 1350 to 1750 Ma-old Tideridjaouine alkaline to peralkaline rhyolite-granite complex. In central Hoggar to the east, the thin, marble-bearing Aleksod Formation, which unconformably overlies Eburnean (about 1.9 Ga) basement (Caby, Bertrand, and Black, 1981), is considered to bear the imprint of the Kibaran "orogeny" (about 1.0 Ga) as obtained on metasedimentary components (Bertrand and Caby, 1978).

Late Proterozoic-Phanerozoic sedimentation upon the West African Shield proper, stable since Eburnean time (about 1.9 Ga), took the form of basin fill upon an essentially bevelled Archean-Eburnean basement. The central part and eastern margin of the shield were gently down-warped to form the vast central Taoudeni Basin and much smaller Volta basin to the southeast (Bertrand-Safarti, Fabre, and Moussine-Pouchkine, 1977; Bonhomme and Bertrand-Safarti, 1982; Black, 1980; Bronner, Roussel, and Trompette, 1980). The resulting sandstone-stromatolitic carbonate cratonic cover, up to 4 km thick (Atar el Hank Group), includes the 700 m-thick Série Carbonatée of age 850 to 700 Ma. This cratonic cover is characterized by deposition on a remarkably flat surface, wide extent, great homogeneity of facies, and average subsidence rate of 5 m/Ma. Specifically before 650 Ma, the subsidence rate was extremely low, 3 to 4 m/Ma; thereafter it rose to 15 m/Ma, a normal value for large cratonic basins (Bronner, Roussel, and Trompette, 1980). The earlier dominantly sandstone phase of Taoudeni and Volta sedimentation (Supergroup 1) (Affaton, Sougy, and Trompette, 1980) began about 1035 Ma ago based on dates of 1020 Ma in Senegal, 1025 Ma in Mali, and 1035 Ma in Algerian Hoggar (Clauer, 1976) and ended with the pre-tillite unconformity dated at 650 to 620 Ma (Katangan Unconformity). In both basins, Supergroup 1 is unconformably overlain by Supergroup 2 of Vendian-Cambrian age containing a triad continental tillite barite-limestone-silexite association at the base. In the Volta Basin Supergroup 2 is overlain by the Obosum Group, a continental red detritic sequence.

The Taoudeni Basin has a simple structure with the youngest formations in the center. Sedimentation was typically regressive. The lithostratigraphic succession is divided into 4 discordant subgroups. Seismic-based isobaths of the basement show a mean thickness of about

1500 m of sediments in the central part of the basin. However, a local north-northeast-oriented trough in the northwest part, a probable aulacogen and the major element of the deep structure of Taoudeni Basin, contains about 4000 m thickness of sediments (Bronner, Roussel, and Trompette, 1980).

In the Pharusian Belt (Pharusides) of the Touareg Shield to the east the "Série a Stromatolites" (Pharusian 1) about 3 km thick and dated at 1070 to 793 Ma (Clauer and others, 1982) is composed of fluviatile quartzite and conglomerate at the base overlain by alternating stromatolitic carbonates with siltstone and quartzite more abundant at the top. It unconformably overlies either Tassendjanet (about 2.0 Ga) basement or Ahnet (pre 1750-1350 Ma) strata. The Série a Stromatolites is correlated in detail with the Atar Group of the West African Shield. Abundant 800 Ma-old (Clauer, 1976) (that is post Série a Stromatolites) basic to ultrabasic sills, laccoliths, and stocks constitute more than 70 percent of all Pharuside basement to still younger Late Proterozoic Série Verte (Pharusian 2) (690-620 Ma) calc-alkalic volcanoclastic deposits (see below). These are, in turn, overlain by Vendian-Cambrian sediments (Série Pourprée).

The remarkable similarity of distinctive, laterally continuous sedimentary facies and stromatolite associations in both the West African Shield (Atar Group including Série Carbonatée) and the adjoining Pharusian Belt (Série a Stromatolites) of the Touareg Shield to the east strongly indicates the presence of a single greater paleo-West African craton (that is West African-Touareg) from at least 1035 to 800 Ma ago (Bertrand and Caby, 1978) which thereafter rifted and drifted apart to form the separate West African Craton and Touareg Shield (see below).

Significantly, to the south in the Gourma region of the eastern Taoudeni Basin, a major change in sedimentation occurred 800 Ma ago due to rifting along the eastern margin of the West African Shield (Caby, Bertrand, and Black, 1981; Black, 1980). The Gourma embayment is thought to be a failed arm which evolved as an aulacogen. It contains terrigenous clastics and carbonates more than 8-km thick, ending with prograde continental clastic deposits. This represents an underlying sequence deposited on a passive continental margin on the eastern edge of the West African Shield and a post-collisional overlying foreland trough.

Furthermore, in the Pharusian Belt, the Série Verte (Pharusian II = Tiririne Formation), dated at 690 to 630 Ma (Clauer and others, 1982), is a 6 km-thick island arc-type andesite-rhyodacite volcanoclastic-intrusive-flysch association. It and still younger andesite-dacite flows and pyroclastics (Toudart, Gara Akefou, and Oumassene) provide typical calc-alkalic geochemical patterns (Caby, Bertrand, and Black, 1981; Chikhaoui, Dupuy, and Dostal, 1980). The Bou Azzer ophiolite, located 1000 km northwest on strike in the Pan-African segment of the Anti-Atlas belt, Morocco, and dated at 788 ± 9 Ma (Leblanc, 1981) was obducted onto the West African Shield about 685 Ma ago (Leblanc, 1976, 1981). These data thereby mark the approximate dates of West African-Touareg ocean-floor opening (about 790 Ma) and closing (about 690 Ma) respectively.

It is noted in passing that the Bou Azzer ophiolite (about 790 Ma) is coeval with early ophiolites in Newfoundland of the North Atlantic domain (Strong, 1979) thereby implying widespread Late Proterozoic (about 800 Ma) plate movements (see table 4).

Post-collisional volcanic-clastic deposits (Série Pourprée), up to 6 km thick and dated at 620 to 470 Ma (Clauer and others, 1982), crop out in north-trending graben and residual basins in the Pharusian Belt. They include a basal group of tillites, carbonates, jaspers, and felsic volcanic-ash, a middle group of continental red arkosic sandstone and conglomerate, and an upper group, exceeding 1.5 km thick, of varied clay, tillite, and graywacke.

In both the Taoudeni and Volta basins horizontal early Paleozoic sandstones are overlain by Ordovician tillites (Affaton, Sougy, and Trompette, 1980). In the Volta Basin this includes the Obosum Group, interpreted as molasse of the Pan-African belt to the east. Fossiliferous Silurian, Devonian, and Carboniferous strata are present to the east in Senegal and Mauretania. Both basins were mildly warped, undulated, and folded during the Hercynian Orogeny (0.4-0.2 Ga) which also resulted in the emplacement of mafic dikes, sills, and laccoliths. Continental deposits of Cretaceous to Recent age are widespread. In addition, two Phanerozoic basins lie respectively to the west (Reggane) and south (Iullemmeden) of the Touareg shield, both within the area of the designated Precambrian composite (table 2, column 4).

In the Benin Nigeria Shield (Turner, 1971; Grant, 1978) to the south, the pre-Mesozoic basement complex largely comprises banded gneiss, migmatite, amphibolite, quartzite, and marble. These yield common Pan-African (600-500 Ma) ages. However, local whole rock Pb ages of 2750 Ma probably reflect the presence of Archean basement. Indeed the bulk of the contained metasedimentary rocks are probably early Proterozoic deposits. In fact, Eburnean age (about 1.9 Ga) rocks could make up most of the basement of Nigeria (Grant, 1978). Amphibolite facies prevails with local granulite and greenschist facies. To the west, rocks of the Benin Nigeria Shield are thrust westward onto the passive margin of the West African Craton to form a possible suture zone. Here the pre-existing ocean opening was either very small, or the oceanic crust and active margin were entirely subducted (Black, 1980). The central part of Nigeria, locus of the mid-Jurassic (160 Ma) tin-bearing alkaline ring-complexes, is underlain by abundant Pan-African granitoids dated at 700 to 500 Ma (Black, 1980).

Central African Shield

The central African Shield (Cahen, Delhal, and Lavreau, 1976; Cahen and Snelling, 1966; Cahen and Lepersonne, 1967; Kroner, 1977a, b; Clifford, 1968; Tanner, 1973; Cahen and others, 1984), as here defined, is a large sub-circular mass 3000 km in diameter and 5,711,000 km² in area (figs. 2, 3; tables 1 and 2, column 2). This mass was originally defined as the Congo Craton (Kennedy, 1964; Clifford, 1970). However, Cahen (1982) applied the term "Congo Craton" to the same mass less the Tan-

TABLE 4

Proposed Africa-America “Early Extension-Later Compression” Cycles¹. Following Gastil (1960) a cyclic pattern about 400 Ma-long is used back to 2200 Ma. Cycles I-III are reliably classified as Wilson (open-close) Cycles. Cycles IV-VI are products of uncertain tectonic process; they possess some but not all of the Wilson Cycle characteristics, hence may be products of a predecessor tectonic process not necessarily involving full-scale intercratonic active-margined (subduction zones) oceanic crust

Extension (ocean)	Compression (supercontinent)	Date ² (Ma)	Cycle (Duration in Ma)	Concurrent Events ³	Main References
Atlantic		0 200	I Current (+200)	Arctic Ocean-Baffin Bay, Sverdrup Basin; Benue and Amazon rifts	Bullard, Everett, and Smith, 1965; Kerr, 1980
	Pangea	400	II Appalachian (400)	Appalachian-Hercynian Caledonian events	Dietz and Holden, 1970
Iapetus		600		Hudson Bay, Franklinian Basin, Volta Basin	Williams, 1980; Roberts and Gale, 1978; Kerr, 1980
	Proto-Pangea I (Pangea E) ⁴	800	III Pan African (400)	Pan African thermotectonic events	Morel and Irving, 1978; Kroner, 1980
Proto-Iapetus I		1000		W. Africa-Hoggar opening; North American rifts and passive continental margins	Leblanc, 1981; Strong, 1979; Rast et al., 1976; Roberts and Gale, 1978; Stewart, 1976

	Proto-Pangea 2 ⁵	1200	IV Grenvillian (400)	Taoudeni and Congo basins subsidence; Grenvillian- Kibaran events	Choubert and Faure-Muret, 1971; Irving and McGlynn, 1976; Dunlop et al., 1980
Proto-Iapetus 2		1400		Grenville and Midcontinent rifting; Borden Basin; Poseidon opening; Kibaran and Hoggar rifting	Irving and McGlynn, 1976; Irving et al., 1974; Jackson and Iannelli, 1981; Bertrand and Caby, 1978; Black 1980
	Proto-Pangea 3 ⁵	1800	V Elsonian (400)	Anorogenic events; Epi- and peri-cratonic basins	Van Schmus, 1981; Raemaker, 1981; Fraser et al., 1970; Black, 1980
Animikean		2200	VI Hudsonian (400)	Eburnean events, Circum- Superior rifting; Wopmay aulacogen; Huronian, Birimian and Ubendian belts	Baragar and Scoates, 1981; Hoffmau, 1980; Sims et al., 1981; Bessoles, 1977; Cahen, Delhal, and Lavreau, 1976

¹ I-III are Wilson (open-close) Cycles; older Cycles of uncertain origin.

² All dates are approximate.

³ See table 1 for pertinent orogenies, mafic dike swarms, et cetera.

⁴ Proto-Pangea I (Pangea E) based on paleomagnetic reconstruction of Morel and Irving (1978).

⁵ Proto-Pangea 2 and 3 combined correspond to Piper's (1976) proposed Pre-Pangea supercontinent.

ganyika (Tanzania) Craton on the east. Kroner (1977b) applied the term "Central African Protoshield" to the same mass less the Kasai craton on the south. Tanner (1973) applied Central African Shield to the eastern craton only. To avoid confusion I call the entire mass the "Central African Shield," the Tanzanian Craton included (fig. 2). Specifically the Central African Shield includes the following 11 Ancient Cratons (Saggerson, 1978), in clockwise succession from the east (fig. 2): Tanzanian, Zambian, Kasai, Angolan, Chaillu, Gabon, Cameroun, Yade, Bomu, Kibalian, and West Nile, together with younger belts and basins. The exposed Central African Shield has comparatively high total Archean content (34 percent) (table 2, column 2). The Shield is characterized by the large central Late Proterozoic-Phanerozoic-filled Congo Basin, 1000 to 1200 km in diameter. Four main orogenies are recorded: Kasaian (Dibayan), 2.7 Ga; Ruzizian (Eburnean), 1.9 Ga; Kibaran-Burundian, 1.1 Ga; and Katangan (Damaran), 0.6 and 0.5 Ga. An older event, Bomu (3.6 Ga), is locally recorded. The Shield is almost entirely bordered by Pan-African (0.6-0.5 Ga) metamorphic belts except for locally rifted Mesozoic oceanic crust of the Atlantic Ocean in the extreme northwest (Gabon-Cameroun).

Archean crust (34 percent) is exposed in the ring of ancient cratons surrounding the Congo Basin (fig. 2). To the south, the Kasai and part of the adjoining Angolan cratons (Cahen, Delhal, and Lavreau, 1976; Cahen and others, 1984) are underlain by (1) extensive tracts of northeast to east-northeast-trending tonalitic migmatite (2.65 Ga) and calc-alkaline granites (2.59 Ga) of the Dibaya Complex, and (2) the Kasai-Lomami Complex of gabbro-norite and charnockitic rocks. Ancient gneiss remnants (Luanyi and Luizian) are dated at 3.49 to 3.38 Ga (Cahen and others, 1984). Extensive granulite grade metamorphism occurred at 2.82 Ga.

To the northwest three ancient cratons (Chaillu, Gabon (Ntem), and Cameroun) include two major units: (1) the "Complexe Calco-magnésien" composed of enderbite, charnockites, and metadolerite (2.90 Ga and older) forms an extensive basement, and (2) the associated Du Chaillu Massif of heterogeneous granitoids (2.66 Ga) (Cahen and others, 1984) associated with gneiss and mica schist (2.7 Ga) (Bonhomme, Gauthier-Lafage, and Weber, 1982), and younger potassic granitoids (2.01 Ga) (Cahen, Delhal, and Lavreau, 1976). Farther east, the Bomu craton includes kyanite-bearing granulite and charnockite dated at 3.00 Ga (Cahen and others, 1984). Model ages of some Bomu mafic gneiss suggest a precursor age of 3400-3700 Ma. This is supported by Pb:Pb data on related Nzangi gneiss which indicate an age of 3417 ± 19 Ma.

Northwest Equatorial Africa is characterized by the extensive west-northwest-trending Kibalian-Toro granitoid-greenstone terrain, also called the Upper Zaire granitoid massif. It is approx 1000 km long and 200 to 300 km wide, with an estimated ratio of granitoid:greenstone = 5:1. Greenstone belts of several different ages are present. Ganguan greenstone belts at the northwest margin of the Kibalian craton, with favored age of 3.2 Ga (Cahen and others, 1984), unconformably overlie Bomu Gneiss

(granulite-charnockite) with suggested precursor age of 3.4 to 3.7 Ga, which is itself intruded by Bondo granitoids (2.56 Ga). The western Kibalian greenstone belts between Uele River and Isiro include varied metavolcanic rocks with economically important BIF. The eastern Kibalian belts of Kilo and Moto include two distinct lithostratigraphic units of different ages. Lower Kibalian strata comprise mafic to intermediate metavolcanic rocks with some BIF. They are intruded by tonalites of an early (first) generation dated at 2.89 Ga. Upper Kibalian strata include substantial andesitic-dacitic metavolcanic rocks, quartzite, and BIF. They are intruded by granitoids of a second generation dated at 2.51 Ga. At Kilo in the east, greenstone belts (Lower Kibalian) are intruded by granitoids dated at 2.72 Ga. Finally, in central Uganda, Kibalian strata of the West Nile (Adida belt) overlie both early-Aruan gneiss (2.68 Ga) and Watian granulites (2.91 Ga) and are themselves affected by the late-Aruan tectono-thermal event dated at 2.55 Ga (Cahen and others, 1984).

In the northern Tanzanian Craton, mainly south of Lake Victoria, the Nyanzian assemblage, at least 7 km thick, comprises older pillowed mafic metavolcanic rocks followed by intermediate to felsic metavolcanic rocks and still younger graywacke, tuffs, slate, and BIF. This assemblage, which may exceed 3.15 Ga in age, is cut by the 2.74 Ga-old Nigori Granite (Cahen, Delhal, and Lavreau, 1976). The unconformably overlying Kavirondian metasedimentary sequence, 1.5 to 3 km thick, comprises a lower division of mainly pelitic rocks with rare grits and volcanic units and an upper division of arkose with pebble bands and conglomerate. Kavirondian strata are affected by a 2.54 Ga-old tectono-thermal event and cut by granitoids dated at 2.47 Ga which provide minimum Kavirondian ages. Correlative metasediments elsewhere in the Central African Shield include the Aruan sequence of West Nile complex and possibly the Luiza Formation of Kasai.

In the southeastern Tanzanian Craton the comparatively small and ill-defined high-grade Dodoman schist belt (Dodson, Bell, and Shackleton, 1973) is composed of quartzite, amphibolite, and gneiss. It was extensively metamorphosed and intruded 2.6 Ga ago.

The entire Central African Shield underwent massive granitoid emplacement and metamorphism 2.8 to 2.6 Ga ago (Kasaian or Dibayan Orogeny).

Early Proterozoic metasupracrustal and associated gneisses (19 per cent) form the prominent southeast-trending Ruzizian (= Ubendian) belt which in the southeast part separates Tanzanian and Zambian cratons. The belt is at least 1000 km long and 250 km wide (fig. 2C). To the north it turns west under the cover rocks of the Congo Basin, and to the southeast it merges with the younger (Pan-African) Mozambique Belt. The Ubendian sequence, about 3.5 km thick, is composed of phyllite, schist, quartzite, and widespread carbonates. Metamorphism is greenschist to amphibolite grade. Extensive reworking of older rocks, including granulite, occurs in the Ubendian part to the southeast. Migmatites are common in the Ruzizian part to the northwest, where they form basement

to younger Kibaran sediments (see below). Younger intrusions include Burundian and Ruzizian granitoids dated at 2.1 to 1.7 Ga (Cahen and Lepersonne, 1967). A second variably dated Precambrian belt (Early-Mid-Late Proterozoic), the Mayumbian, situated 2200 km to the west and paralleling the Atlantic coast from Gabon to Angola, is at least 1000 km long and 80 to 100 km wide (Cahen and others, 1984). The belt is divided into northern Kimezian and southern Zadinian parts (Lavreau, 1982). The Mayumbian Supergroup (Mid-Proterozoic) is similar in composition to the Ubendian but with notable rhyolitic volcanic content in the upper part. Slates, phyllites, and schists in the easterly part of the belt grade westward to amphibolite grade gneiss. Two post-Mayumbian granitoids are dated at 1027 ± 56 Ma and about 1057 Ma (Cahen, 1982). Late Proterozoic (Pan-African) dates are common, especially in the southern part (Saggerson, 1978).

To the northeast, Francevillian sediments of Gabon unconformably overlie Chaillu-Gabon basement. Francevillian deposition occurred some time between 2700 and 2050 Ma (Bonhomme, Gauthier-Lafage, and Weber, 1982). Diagenesis is dated at 2050 to 1870 Ma. The sediments comprise sandstone, conglomerate, shale, dolomite, and tuffs. They are remarkable for their uranium ores with natural nuclear reactions (1930-1980 Ma).

To the north, Lower Proterozoic candidates exist among the "séries intermédiaires" of Cameroun-northwest Zaire (Lavreau, 1982). To the northeast, in northern Zaire and Central Uganda, in the 300 km-long interval between the easternmost Kibalian greenstone belt (Lake Mobutu) and the northernmost Nyanzian-Kavirondian greenstone belt (Lake Victoria), the Lower Proterozoic Ruwenzori fold belt (Tanner, 1973) contains the Buganda-Toro Supergroup which falls in the 2500 to 1850 Ma time range (Cahen and others, 1984). The Buganda group of central Uganda comprises, from bottom to top, quartzite, phyllite, slate, amphibolite, and tuff. The Toro "system" of western Uganda comprises a schist and gneiss component (the Igara schists) and the Toro quartzites.

Additional Early Proterozoic rocks are represented by four polycyclic cratons. To the southeast the Zambian Craton represents an ancient Archean mass characterized by granitic gneiss, granulite, and amphibolite associated with low metamorphic grade greenstone belts, all extensively rejuvenated by Eburnean (Ruzizian) metamorphism (Saggerson, 1978). Post-Eburnean molasse (Plateau) locally overlies the craton. A similar Archean-Eburnean polycyclic block with Early Proterozoic cover rocks is present in Angola to the southwest. In Cameroun Craton to the northwest, and in parts of the Kibalian block to the northeast (fig. 2C), Eburnean and Pan-African rejuvenations are widespread.

Eburnean tectogenesis peaking at about 2.06 Ga (Nyika gneiss) generated vast amounts of granitoid intrusions and belt mobility (Saggerson, 1978). Sn, W, Pb, and Au mineralization is associated with Eburnean (Ruzizian) granitoids.

Mid-Proterozoic rocks (7 percent).—The prominent east-northeast-trending Kibarides (= Burundian = Karagwe-Ankolean in the northeast) and Irumides (= Rehobothian in the southwest) located respectively northwest and southeast of the Zambian Craton, constitute well-defined linear belts each 1000 to 1300 km long and 100 to 500 km wide. The belts represent localized mobility in restricted fault-bounded zones. They are composed of uniform assemblages, each 10 to 15 km thick, of pelitic-arenaceous metasediments, mainly slate, schist, quartzite, marble, and calc-silicates with some volcanic rocks, which were deposited on older, eroded Ruzizian-Ubendian basement. Deposition occurred in the interval 1850 to 1310 Ma (Cahen and Lepersonne, 1967). Early granitoids were emplaced 1300 Ma ago, and late tectonic granitoids about 1190 Ma ago (Saggerson, 1978). Kibaride belts transect the older Ubendian-Ruzizian mobile belt and were, in turn, deformed and metamorphosed (mainly low grade) 1370 to 1310 Ma ago (Cahen, 1982). Sn, Ta, W, and Li mineralization is associated with 1000 Ma-old pegmatites. Some Zn, Pb, and Au deposits are contained in Kibaran strata.

Late Proterozoic rocks (14 percent).—Little folded or unfolded Katangan arenaceous-pelitic-carbonate sedimentary rocks of the Katangan Supergroup and stratigraphic equivalents (West Congolian, Mbuji Mayi, Lindian, Itombwe, and Malagarasian-Bukoban supergroups) crop out around much of the periphery of the Congo Basin (fig. 2C) and appear to underlie virtually all of it. They lie with marked unconformity on older rocks and are unconformably overlain by Phanerozoic cover rocks. Katangan Supergroup, typical of Late Proterozoic sedimentation, started 1200 to 1100 Ma ago (Roan or lower Katangan); this was followed sequentially by the Mwashya sequence at about 1000 Ma, the "Grand conglomerat" and lavas of the Lower Kundelungu at 950 Ma, and the "Petit conglomerat" of the upper Kundelungu at 788 Ma. Katangan accumulation ceased about 602 Ma (Cahen, 1982). Katanga-type sedimentation is notable for the presence of (1) mixtites, possibly glaciogenic; (2) stromatolites (Bonhomme and Bertrand-Sarfati, 1982); (3) uppermost redbeds; and, in Zambia, (4) unusually extensive Cu-Co-U mineral deposits (Copper belt). Based on stromatolite zonation, Mbuji Mayi sediments (formerly Bushimay Supergroup [BII group]) of western Zaire, which correspond to the Roan Group of the Zambian Copper Belt, are correlated by Bertrand-Sarfati (1972) with beds low down in Supergroup I of the Taoudeni Basin of the West African Shield dated at 886 ± 16 Ma or older (Clauer and others, 1982).

As many as six separate metamorphic events of varying intensity are recorded in the Katanga belt (Kroner, 1980). The oldest is at 950 Ma, and the others in the interval 720 to 550 Ma. Volcanic rocks and granitoid intrusions are notably rare in this belt, (Cahen, Delhal, and Lavreau, 1976).

Generally, Pan-African events took place between 750 and 450 Ma ago. They are recorded in nearly half the crystalline rocks of the African continent, being particularly evident in the younger mobile belts circum-

jaacent to the ancient cratons (Kennedy, 1964; Saggerson, 1978). They display a bimodal spread of ages, the first peak at 620 ± 20 Ma and the second at 500 ± 30 Ma. Metamorphism is frequently of intermediate pressure facies type although high pressure deformation is recorded locally, for example, Damaran and Zambian.

Two vertical drill holes in the Congo Basin (Lepersonne, 1977), each about 2000 m deep, penetrated more than 1000 m of flat-lying to gently folded Precambrian fine-grained red arkosic sandstone beneath the Phanerozoic basin fill. Although the sandstone has not been directly dated, it is known to have been involved in the last phase of Precambrian folding dated at 650 to 600 Ma (Lepersonne, 1977) and is presumably Katangan. Also, as noted above, stromatolites in exposed correlative strata provide some additional support for an upper Proterozoic (about 886 Ma) age.

Phanerozoic rocks (26 percent).—A gap of 300 Ma followed the Late Proterozoic (= Vendian)–Cambrian (570 Ma) sedimentation, recommencing at upper Carboniferous (280 Ma). Permian and Triassic strata are also represented, but the bulk of the Congo basin fill represents continental deposits ranging from Jurassic to Quaternary. The most complete Mesozoic sequence is to the north. Three marine incursions are represented by the Kwango, Loia, and Stanleyville series, respectively, separated by slight unconformities (Mestraud, 1971).

Tectonic Summary

The *Canadian Shield* is formed mainly of Archean and Early Proterozoic rocks (about 80 percent). Archean crust is dominated by Superior Province, an unusually large and diverse greenstone- and granulite-rich ancient craton. The broad distribution of small to large Archean cratons and of intercratonic rejuvenated Archean crust indicates clearly that varying stabilized Archean crust extended shield-wide by Kenoran (2.6 Ga) time. Immense quantities of juvenile granitoids — the main cratonizing agent — were added to the crust during this terminal orogeny.

The continuity and preservation of circum-Superior Early Proterozoic belts and basins replete with shelf facies effectively demonstrate that Superior Craton has not been significantly modified in surface dimension or relative vertical position (epeirogenesis) since Aphebian deposition about 2.0 Ga ago. Most Aphebian belts are “early extension-later compression” type with disputed oceanic crust involvement. Some are normal greenstone belts. Others are considered to be ensialic. Belt mobility commonly involved basement-cover mobilization along broad thermal-anatectic fronts with significant cratonward (especially toward Superior Province) tectonic transport. A major result has been to juxtapose widespread catazonal granulites against mesozonal and even epizonal crust. Granulite-upper amphibolite facies rocks are especially common at Churchill Province perimeters, sites of preferential vertical uplift. In brief, horizontal shortening involving crustal telescoping, interleaving and stacking, together with diapiric intrusion of resulting deeper crustal partial melts, has contributed massively to shield-wide cratonization.

Widespread Hudsonian granitoids (about 1.8 Ga) represent mainly rejuvenated Archean crust.

Early Proterozoic crust, now concentrated in the central-northern shield, may at one time have been equally widespread south and southeast of Superior Craton. If so, such crust was largely obliterated by later (Hudsonian, Grenvillian, and Appalachian) tectonic events involving northwestward allochthonous transport and crustal rejuvenations. Many Early Proterozoic belts of the Canadian Shield have crude orthogonal (northeast-northwest) alignments. This may reflect a fundamental, recurrent Early Precambrian shield-wide tectonic pattern. Early Proterozoic (2.1 Ga) aulacogens in the Great Bear region probably document early extension and rifting at the margins of the future shield, a pattern subsequently enhanced shield-wide.

Mid-Proterozoic basins are fault-controlled, redbed sandstone-basalt-filled, epicratonic and pericratonic basins that mark extensive rifting especially at and near present shield margins. Together with extensive diabase dike swarms and local alkaline igneous intrusions, they demonstrate shield-wide extension, again of common orthogonal (northeast-northwest) pattern. In sharp contrast to such tensional phenomena, the northeast-trending Grenville mobile belt, possibly involving early rifting, orogenically peaked at 1.1 to 1.0 Ga, the time of massive cratonward (northwest) and upward tectonic transport and compression ascribed by some to continent-continent collision. The north-northeast-trending Kapuskasing Structural Zone, crossing southcentral Superior Province, also underwent repeated vertical movements and alkaline igneous intrusions. Important coeval (about 1.4 Ga) anorogenic gabbroic to anorthositic, mangeritic, and granitoid (rapakivi)-rhyolitic associations occurred in the eastern shield and in buried midcontinent extensions to the southwest (Muehlberger, 1980; Muehlberger, Denison, and Lidiak, 1967; van Schmus and Bickford, 1981; Anderson, 1983). The gross outline of the present shield was firmly established as a result of widespread late-Helikian (1.1 Ga) aulacogen-bearing pericratonic rifting. This included the opening of Poseidon, the proto-Arctic ocean, to the north.

Late Proterozoic activity (1.0-0.6 Ga), mainly confined to the north, involved rift-controlled pericratonic basins and geosynclines, major diabase dike swarms (Franklin), and extensive northwest-trending normal faults (Baffin Island vicinity), all products of widespread episodic crustal extensions. Iapetus, the proto-Atlantic ocean, began to open about this time (580 Ma).

In mid-Ordovician time (450 Ma ago), coincident with the Taconic orogeny marking Iapetus arc collision to the east, the central shield subsided, mainly along northwest-northeast faults, to form Hudson Bay and Foxe Basin. Michigan and Illinois Basins subsided coincidentally to the south. Mesozoic (about 150 Ma) pericratonic rifting initiated opening of the Arctic Ocean and Baffin Bay. Pleistocene ice loading with consequent crustal downwarping accentuated Hudson Bay which is currently isostatically rebounding.

Clearly, the growing and stabilizing Precambrian shield underwent repeated local epeirogenic uplift and erosion in response to orogenic events and sealevel changes. Specifically, (A) parts of Superior Province were rapidly uplifted and eroded in post-Kenoran (2.6 Ga) time sufficient to (1) produce voluminous Early Proterozoic orthoquartzite fill (for example, Huronian at 2.4 Ga) and (2) provide extensive Archean granulite-rich basement to Early Proterozoic (about 2.0 Ga) cover; local post-Kenoran uplift of 20 km is estimated in Baffin Island; (B) rapid post-Hudsonian (1.8 Ga) uplift of Churchill Province crust was required both to fill 1.7 to 1.5 Ga-old fault-bounded grabens and major epicratonic basins (for example, Athabasca) and to expose extensive granulite-rich basement to Mid-Proterozoic cover, this requiring an estimated 30 km uplift in Baffin Island alone; and (C) rapid post-Grenvillian (1.1 Ga) uplift and erosion to an estimated depth of 15 km exposed widespread catazonal (granulite) Grenville crust. The overall result of these and other related activities was shield-wide exposure of dominant (about 90 percent) granitoid crust comprising about 25 percent granulite and 60 percent amphibolite metamorphic facies. This is considered to mark a net shield uplift and erosion of 10 to 15 km. Much of the resulting Early Proterozoic (2.5-1.7 Ga) detritus is contained in intracratonic mobile belts and epicratonic basins, and much of the mid-late-Proterozoic detritus is disposed in off-shield pericontinental fold belts (Appalachian, Cordilleran, Innuitian).

The *West African Shield* contains two exposed early Precambrian cratons. Each is composed of (1) western granulite-rich Archean part and (2) central-eastern amphibolite-greenschist-rich Early Proterozoic part, the latter comprising rejuvenated Archean basement and north-trending Early Proterozoic supracrustal-rich belts (Birimian and Aftout). Allochthonous sheets have been tectonically thrust onto the shield at the northern, eastern, southwestern, and western margins. The entire shield was essentially cratonized by Eburnean time (1.9 Ga). Eburnean granitoids are abundant. Mid-Proterozoic rocks (1.9-1.1 Ga) are virtually absent (La Grande Lacune). During this 700 to 900 Ma hiatus the entire shield was uplifted, eroded, and peneplaned. This broad flat surface then subsided at an average rate of 5 m/Ma to form the very large central Taoudeni Basin and much smaller Volta Basin to the southeast. Basin sedimentation was initiated 1035 Ma ago (Mid-Late Proterozoic boundary) and has continued intermittently to the present. As a result, Late Proterozoic-Phanerozoic cover is the dominant (63 percent) West Africa Shield component (table 2, column 3), as is even more so the case (73 percent) in the West African-Touareg-Benin Nigeria composite (table 2, column 4).

The Touareg and Benin Nigeria shields to the east both have similar Archean-Early Proterozoic histories. Whether they formed unbroken eastward extensions of the West African Shield during the early Precambrian (> 1.7 Ga) is still unproved but plausible. However, the remarkable similarity of stromatolitic Late Proterozoic sedimentary facies in both the West African and adjoining Touareg shields strongly suggests that the two cratons formed a single unbroken craton from at least 1035 to 800 Ma

ago, that is, during early Taoudeni Basin filling. Thereafter, the association of major Pharusian (Late Proterozoic) calc-alkaline igneous rocks, aulacogen sediments (Gourma), and allochthonous ophiolites demonstrate that the two were (1) temporarily (788-685 Ma) separated by oceanic crust of undetermined width, then (2) reunited to represent contiguity about 685 Ma ago. The Touareg and Benin Nigeria shields, but not the much larger West African Shield, were widely affected by Pan-African deformation, metamorphism, and magmatism (0.6-0.5 Ga).

The *Central African Shield*, with comparatively high Archean content (34 percent), is dominated by a ring of early Precambrian ("ancient") cratons which practically circumscribe the Late Proterozoic-Phanerozoic-filled Congo Basin. Archean crust is estimated to have underlain this entire shield area. Both the exposed south-southeast-trending Ruzizian-Ubendian and Mayumbian fold belts, the latter also including younger Proterozoic ages, together with the Buganda-Toro belt to the northeast, illustrate the prevailing pattern of Early Proterozoic intracratonic mobility. Clearly the whole area of the Central African Shield formed a single post-Eburnean (1.9 Ga) craton comprising preserved Archean cratons with circumjacent Early Proterozoic mobile belts. Mid-Proterozoic (1.9-1.1 Ga) mobility, as known, is restricted to the fault-induced north-northeast-trending Kibarides and Irumides in the southeastern quarter. Otherwise the shield behaved as a stable unit. Uplift and erosion during shield history have exposed granulite-rich terrains especially in north-western and southern parts.

By the start of Late Proterozoic time (1025 Ma) the central part of the shield had subsided to form the initial Congo Basin. With continued though intermittent subsidence, Late Proterozoic epicratonic arenaceous-pelitic-carbonate rocks came to underlie virtually all the Basin and at least the adjoining pericratonic Katanga-Damara belt to the south. The succeeding Pan-African tectonism affected practically all rocks circumjacent to the shield but not the shield itself which was practically unaffected.

Shield comparisons.—The three ring-shields (columns 1, 2, and 4 in table 2) have remarkably similar (1) size, (2) lithology, and (3) growth patterns. Common to all three are three main, almost synchronous granitoid-emplacing shield-building orogenies (table 1) which serve to set off four characteristic tectonic phases: (1) *Archean phase* (3.8-2.6 Ga) featuring granitoid-greenstone and high-grade gneiss terranes distributed in discrete ancient cratons; (2) *Early Proterozoic phase* (2.6-1.8 Ga) characterized by circum-cratonic "early extension-terminal compression"-type mobile belts involving massive foreland tectonic transport with accompanying granitoid emplacement, to weld together massive shield-wide early Precambrian basement to later Precambrian-Phanerozoic cover; (3) *Mid-Proterozoic phase* (1.8-1.0 Ga) of comparative shield stability yet widespread local crustal extension in the form of rift-induced epi- and peri-cratonic basins, mafic dike swarms, and local marginal mobile belts, for example, Grenville, Pharuside, Kibaride; and (4) *Late Proterozoic*

Phanerozoic phase (1.0-0 Ga) of enhanced crustal extension characterized by epi- and peri-cratonic rifting culminating in central basin collapse.

Thus, within the context of the continent-wide enclosing Precambrian platforms, common ring-shield development may be summarized in three tectonic stages: (1) early Precambrian *constructional stage* (> 3.8-1.8 Ga) involving widespread recurrent dominantly compressional tectonics and granitoid emplacement leading to large (3000 km diam), massive, buoyant subcircular shields, (2) a moderately extensional *intermediate stage* (1.8-1.0 Ga) of variable "turn-around" tectonics characterized, however, by prevailing shield uplift and stability leading to (3) late Precambrian-Phanerozoic *fragmentation stage* (1.0-0 Ga) of overall extensional tectonics and intershield belt mobility culminating in central basin collapse and basement cover. This + 3.8 Ga-long crustal progression featured a concurrent about 400 Ma-long cyclic pattern extending back to at least 2600 Ma expressed in repeated "early extension-later compression" (latterly Wilson) cycles (see below).

Despite local variations, most Precambrian shields and platforms of the world display this generalized stage-by-stage development, the early Precambrian constructional stage producing common Precambrian basement to late Precambrian-Phanerozoic cover with local intershield belt mobility. The common ring-shield diameter of about 3000 km which is shared by numerous continental platforms (for example, Siberian) apparently expresses optimum tectonic growth, prelude to craton instability and collapse.

On a global scale the large size and tectonic coherence of the three ring-shields express unusual late stage tectonic stability (see also Pretorius, 1973). This stability is ascribed to deep, episodically-developed sub-shield tectospheric roots (see below). This common stage-by-stage three-dimensional shield development is reasonably attributed to global first-order thermotectonic events affecting a cooling (McKenzie and Weiss, 1980), essentially uniform volume (McElhinny, Taylor, and Stevenson, 1978), lithosphere-thickening Earth.

Thus stage-by-stage crustal evolution is attributed to evolving lithosphere. Early constructional stage reflects comparatively ductile lithosphere expressed either in common ensialic belts or in small restricted plate motions (Wynne-Edwards, 1976), whereas late fragmentation stage reflects brittle plate-tectonic lithosphere with large-scale Wilson Cycle plate dispersion-reassembly involving substantial subduction-provoking intercontinental oceanic crust. The intervening "turn-around" stage, in turn, correlates with the proposed 1.8 to 1.0 Ga-old (Mid-Proterozoic) pre-Pangaeian supercontinent (Piper, 1976), a time of comparative continental stability. Mid-Proterozoic supercontinent stability finds support in Australia (Veevers and McElhinny, 1976), Laurasia (Irving and McGlynn, 1976; McGlynn and Irving, 1981), and Gondwanaland (Donnet, 1976). Other supporting data, of varying applicability, include (1) geochronologic provinces (Hurley, 1970), (2) accordant tectonic patterns (Engel and Kelm, 1972), (3) converging tectonic lineaments (Dearnley, 1966), (4)

Early Proterozoic banded iron-formations (Goodwin, 1973), and (5) Mid-Proterozoic anorthosite belts (Hertz, 1969). In brief, available data support without proving the presence of a Mid-Proterozoic Pangaea-type supercontinent of exceptional interior tectonic stability. However, Mid-Proterozoic stability does not deny the presence of either normal pericratonic mobile belts or restricted internal belts, the latter marked by collisional zones products of restricted plate motions.

The Canadian Shield proper is mainly surrounded by Phanerozoic cover upon Precambrian basement, the latter rich in Early-Mid-Proterozoic mobile belts (to 1.0 Ga). The two African shields, in contrast, are mainly transitional to circumjacent Pan-African (0.6-0.5 Ga) mobile belts, a Pangeaic mobility (Hurley, 1974) which is restricted in eastern North America to the somewhat younger (0.4-0.3 Ga) Appalachian belt (fig. 4). Clearly the two African shields themselves underwent a withdrawal from further active igneous and metamorphic "rejuvenation" following the earlier Kibaran-Burundian Orogeny about 1.0 Ga ago. Concurrently, practically the entire North American platform, the Canadian Shield included, underwent a similar withdrawal following the Grenvillian Orogeny (1.1-1.0 Ga). Thus although all three ring-shields share a common early Precambrian-rich composition (table 2), the rejuvenation patterns vary by shield and platform, a variation tentatively attributed to differing tectospheric roots (see below).

The ring-shields provide no convincing evidence of primary "continental" accretion (Wilson, 1949). Rather the evidence favors a "jigsaw" pattern of craton growth, aggregation, and stabilization with, however, repeated patterns of centripetal (that is cratonward) tectonic transport such that hinterland catazonal crust now typically adjoins foreland mesozonal-epizonal crust. This, then, constitutes a type of peripheral tectonic accretion. Certain crudely concentric patterns in Early-Mid-Precambrian crust of the southern Canadian Shield, tentatively ascribed to meteorite impact (Klasner and Schulz, 1982), are more reasonably attributed to centripetal tectonic transport, as are crudely concentric shield- and platform-wide North American patterns, younger dates outward (Gastil, 1960).

Pre-1.0 Ga mobile belts are of generally uncertain tectonic origin. Some contain calc-alkaline igneous rocks which are typical modern subduction zone associates and even aulacogen-type sedimentary sequences, all characteristic of Wilson Cycle plate tectonics. Some or all of these belts may, indeed, be Wilson Cycle products. However, in the absence of convincing evidence of oceanic crust (ophiolites, et cetera) and with regard to some possible non-subduction-related calc-alkaline-producing predecessor process, I prefer to call them "early extension-later compression" cycles. However, the presence of 0.8 Ga-old ophiolites in both the West African and Canadian (Newfoundland) shields marks full-scale Wilson Cycle plate tectonics in Late Proterozoic time at least (Kroner, 1981; Goodwin, 1981b).

ORIENTATION OF PAN-AFRICAN BELTS TO TIE-LINE

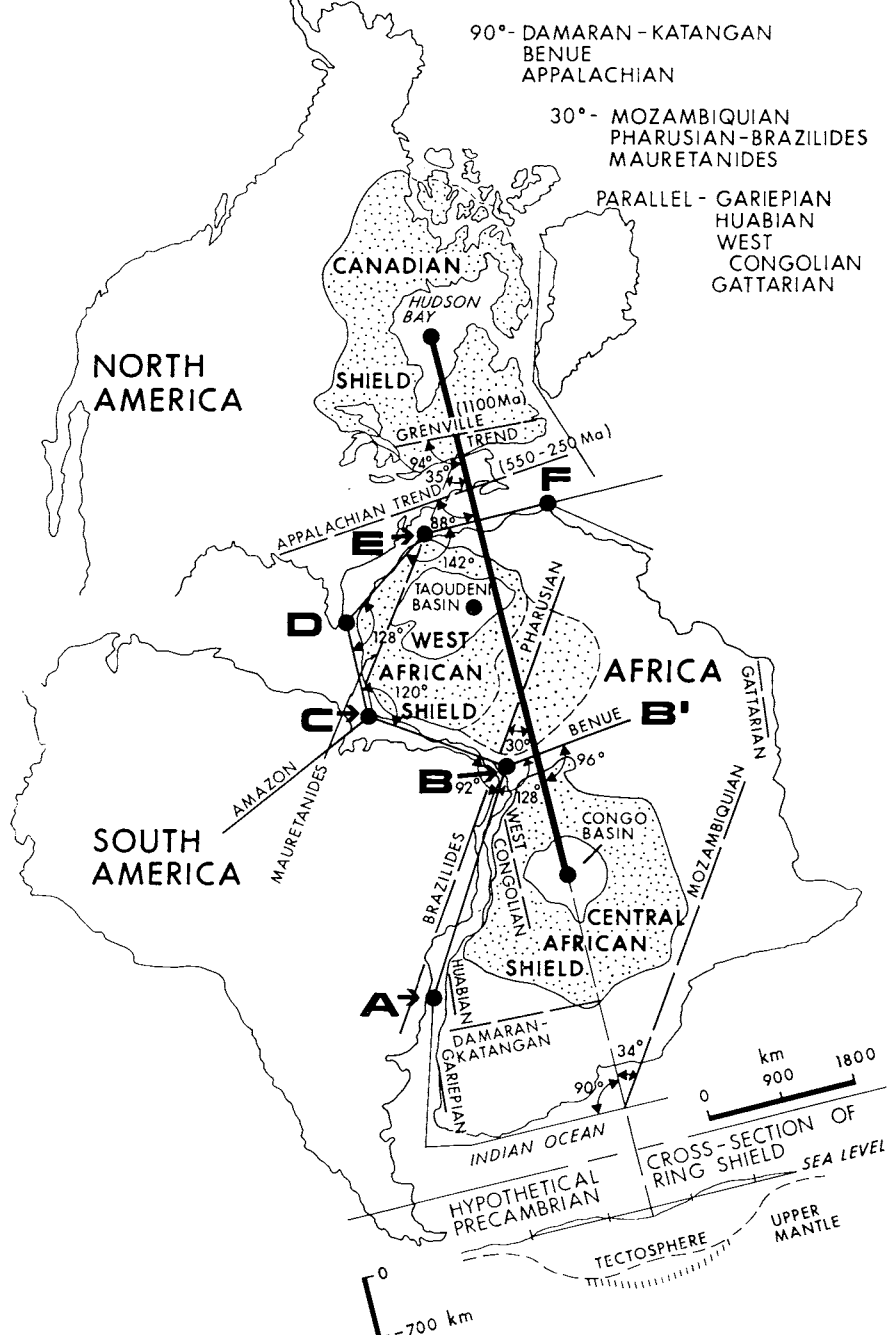


Fig. 4. The Canadian-West African-Central African ring-shield alignment (Hudson Bay-Taoudeni Basin-Congo Basin tie-line) on the Bullard, Everett, and Smith (1965) reconstruction of 200 Ma ago. The basin tie-line is shown relative to (1) selected Mesozoic spreading ridges (A-B-C-D-E-F) of the future Atlantic Ocean and (2) the orientation of Pangeaic (0.7-0.3 Ga) and earlier fold belts (Hurley, 1974), in and around the shields. Also shown is a hypothetical cross section of a Precambrian ring-shield with deep tectospheric root zone including possible deeper zone of chemical or phase-change sub-root down to the 700 km-deep upper mantle boundary and deeper.

The Canadian and two African ring-shields have generally similar yet partly distinctive metallogenic patterns (table 1). Archean Fe and Au deposits are common to all, but Archean Cu-Zn sulfide deposits and Cr, Ni, and Mn deposits respectively characterize the Canadian and African shields. Early-Mid-Proterozoic patterns differ by shield with (1) U, Ni-Cu-Pt, and Ti deposits and (2) diamonds and Sn-W deposits respectively characterizing the Canadian and African shields. Late Proterozoic (about 1.1 Ga) Cu mineralization is common to all three shields.

Considering Precambrian content only, the three ring-shields, each about 6×10^6 km² in area, are equally divided between Archean (D), Early Proterozoic (C), and Mid-Late Proterozoic (B + A) rocks (table 2, columns 5, 6). Thus, Precambrian D:C:B:A = 6:6:2:4. Early Precambrian (C + D) rocks (66 percent) comprise common basement, and Late Precambrian (B + A) rocks (34 percent) common cover; thus basement:cover = 2:1. Considering Precambrian proportions by shield, Archean content is highest in the Central African Shield (11 Ancient Cratons), Early and Mid-Proterozoic in the Canadian Shield (Churchill-Southern mobile belts and Grenville Province respectively), and Late Proterozoic content in the West African Shield (basin fill and Pharosides).

Considering total (Precambrian and Phanerozoic) average composition in the three ring-shields (table 2, columns 5, 6), Precambrian:Phanerozoic = 2:1; and considered by shield, the West African (columns 3, 4) has almost twice the Phanerozoic content of the other two (columns 1, 2), this reflecting widespread Eocambrian-Phanerozoic basin subsidence in West Africa. The West African Shield, in this regard (that is, high Phanerozoic content), more closely resembles many other Precambrian continental platforms, for example Russian, Siberian, and even Australian (fig. 1) which underwent extensive Mid-Late Proterozoic interior collapse to reach their present status of peripheral shield fragments about large Late Precambrian-Phanerozoic-covered interior basements. Thus the three ring-shields, notably Canadian and Central African, differ from most Precambrian continental platforms in their high degree of late stage tectonic stability, a stability tentatively attributed to the presence of unusually deep tectospheric roots (see below).

The characteristic central basins of the three ring-shields mark Late Proterozoic-Phanerozoic tectonic events. The main African basins (Taoudeni and Congo) were subsiding 1050 Ma ago, concurrent with the Kibaran-Burandian (and Grenvillian) orogeny (table 1). In contrast Hudson Bay of the Canadian Shield only began subsiding 450 Ma ago, coincident with the first arc collision in Iapetus, the proto-Atlantic Ocean (see below). The reason for the 600 Ma Hudson Bay "delay" (that is, 1050-450 Ma) is obscure, as are, indeed, the causes of epicratonic basin subsidence in general. In this regard, of the proposed mechanisms for basin development (Turcotte, 1980), thermal subsidence and tectonic thinning are considered most applicable (Goodwin, 1974). If so, the 600 Ma Hudson Bay "delay" may simply reflect a corresponding North American delay relative to Africa in sub-shield thermal subsidence.

Alternatively but less likely, a potential central collapse-inducing stress field in the Canadian Shield 1.1 Ga ago may have been relieved by massive pericratonic rifting (Lake Superior-Kapuskasing-Baffin Island-Poseidon Ocean), a peripheral stress relief not then available to African crust, on the brink of widespread Pan-African circum-shield thermotectonism.

RING-SHIELD ALIGNMENT AND OSCILLATION

Present distribution.—The three ring-shields, represented by the Hudson Bay-Taoudeni Basin-Congo Basin tie-line, are not only straight-aligned but form part of a great circle (figs. 1, 3). This great circle may be neatly divided into three equal sectors each subtending 120° at the Earth's center (fig. 1). Sector I which crosses eastern North America, the North Atlantic Ocean, and western and central Africa includes the three ring-shields. It has experienced repeated extensional plate spreading represented today by the 5000 km-wide North Atlantic Ocean, product of the past 200 Ma. Sector II transects the Southwest Indian Ocean Ridge, paralleling the Southeast Indian Ocean Ridge to New Zealand; three present shields (Indian, West Australian, and East Antarctic), remnants of a larger, pre-existing Precambrian supercontinent, have been laterally dispersed relative to this line during the same 200 Ma period. Sector III neatly bisects the Pacific Ocean Basin passing close to Hawaii (hot-spot) and transecting at right angles the Cordilleran Belt near Vancouver. This curious ring-shield-embracing great circle, then, bisects in its course the North American, African, and Pacific plates. Furthermore it touches the Atlantic-African and central Pacific residual geoid highs, both currently centered on the equator (Anderson, 1982). Conceivably this great circle reflects some as yet undisclosed fundamental control in crustal development.

Mesozoic reconstruction.—A salient feature of the Bullard, Everett, and Smith (1965) reconstruction and subsequent reconstructions of the continents about the Atlantic Ocean (fig. 3) 200 Ma ago is that the three ring-shields are similarly straight-aligned. In the figure the rift pattern in existence 200 Ma ago, which thereafter gave rise to the Atlantic Ocean, is illustrated in highly generalized form by the trends of continental overlap and gap in the reconstruction. This pattern conforms to the net trends of Mesozoic-Cenozoic spreading ridges and transform faults that have controlled the Atlantic opening. These trends are aligned at or close to 0° , 30° , 60° , or 90° to the designated ring-shield tie-line (fig. 4). The main segments and their orientation to this tie-line are as follows: segment A-B (Walvis-Benue)- 30° , B-C (Benue-Amazon)- 60° , C-D (Caribbean)- 0° , D-E (Carolinas)- 58° , and E-F (Moroccan)- 90° .

This rift pattern can be reasonably attributed to a family of triple junctions (Burke and Dewey, 1973; Dewey and Burke, 1974; Burke and Wilson, 1976), with spreading arms dispersed at 120° to each other and with one arm of each triple junction oriented either parallel to or perpendicular to the ring-shield tie-line. Following Wilson (1968a), the pattern of rifting is such that the rifts approach but do not transect

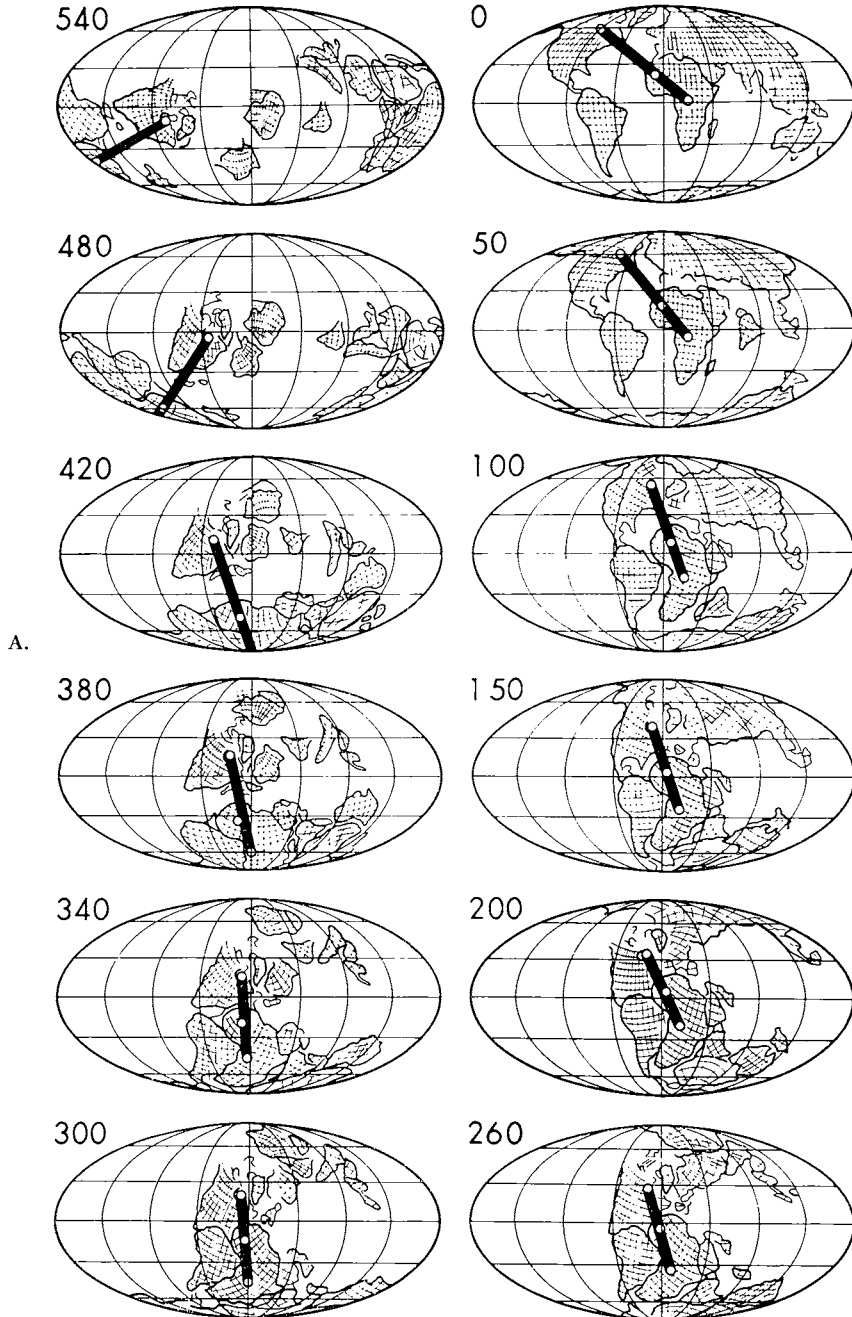
the main mass of the three ring-shields. As a result gross shield integrity has been preserved despite (1) the presence of marginal slices of both African shields across the Atlantic in Brazil and (2) the eastward drift of Greenland from North America. Two different types of intershield rifts are represented: (1) segment E-F (Moroccan) represents the major operating North Atlantic spreading ridge, and (2) segment B-B¹ (Benue) represents a failed arm or aulacogen. It is emphasized that the three-shield alignment has been maintained during Atlantic and predecessor openings because these particular intershield spreading arms are perpendicular to the ring-shield tie-line. As a result, Wilson Cycle motions have returned the continents across the rifts to their starting points or nearby.

It is postulated that the Mesozoic rifting about the future Atlantic was either controlled or significantly influenced by the Precambrian ring-shields themselves such that consequent rifting molded itself about these shields without substantially reducing their size or altering their alignment. Deep sub-shield tectospheric roots are considered to have been the cause (see below).

Other possibly significant tectonic orientations are apparent (fig. 4). Many Pangeaic fold-belts, mainly Pan-African (0.6-0.5 Ga), are systematically oriented relative to the ring-shield tie-line. Thus, the Damaran-Katangan, Benue, and Appalachian (0.5-0.3 Ga) trend at 90° as does the still older (1.1 Ga) Grenville trend (fig. 3). The Mozambiquian, Pharusian-Brazilide, and Mauretanic (0.5-0.3 Ga) trend at 30°; the Gariepiian, Huabian, West Congolian, and Gattarian (not illustrated) belts are parallel or sub-parallel to the tie-line as are the still older (1.9 Ga) Ruzizian-Ubendian, Mayumbian, and Labrador (see fig. 3) belts. In brief, Precambrian fold belt orientation may reflect some form of sustained deep-seated tectonic control during shield growth.

A second fundamental global Precambrian shield alignment crosses Eurasia, West Africa, and South America in the Mesozoic reconstruction (fig. 3). It is oriented at 60° to the designated North American-African shield alignment. Significantly, the South American-Eurasian alignment also crosses Taoudeni Basin of the West African Shield, this basin being common to both alignments. Eastward from Taoudeni Basin the South American-Eurasian alignment successively crosses the Iberian Precambrian mass, Massif Central, Bohemian Massif, and the Russian and Siberian Platforms (fig. 1). Westward from Taoudeni Basin the same alignment follows the Amazon rift, a basin-bearing collapse feature separating the Guyanan and Brazilian shields of the South American platform. These two designated global alignments between them transect all the Precambrian shields and platforms as assembled 200 Ma ago thereby providing the characteristic U- or arrowhead-shaped supercontinent of Pangea (Dietz and Holden, 1970). Of the two designated transecting alignments, the North American-African alignment was predominant in that during the Atlantic opening it was maintained (see below) at the expense of the South American-Eurasian alignment which was broken as a result of rotation of South America away from Africa.

Fig. 5

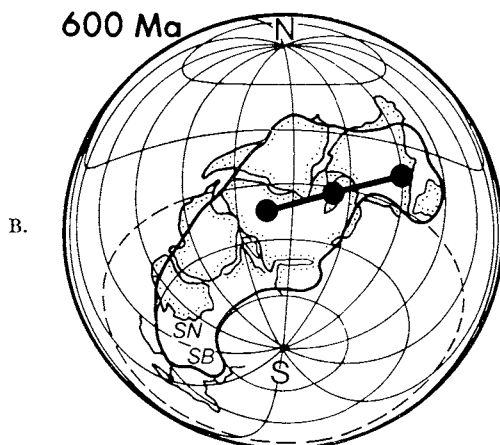


(A) A 12-frame continental reconstruction at approx 50 Ma intervals for the Phanerozoic after Scotese (1979). The three ring-shield basins and tie-line are illustrated where possible. They collectively illustrate (1) maintenance of triple-shield alignment during oscillatory ocean-floor spreading, and (2) progressive global counter-clockwise rotation of the aligned continental fragments during the Phanerozoic. The significant straight alignment is between the line connecting the Center points of the two African shields which shields were joined about 600 Ma ago on the one hand and the center point of the Canadian Shield (Hudson Bay) on the other hand.

The two designated (North American-African and South American-Eurasian) global Precambrian shield alignments (fig. 3) are respectively subparallel to O'Driscoll's (1980) dominant north-northwest-trending Tethyan and subordinate north-northeast-trending Laurasian lineaments. A principal point of contrast, however, is that both O'Driscoll's Tethyan and Laurasian lineaments track out helical patterns on the present globe whereas the designated North American-African Precambrian ring-shield alignment is part of a great circle. However, the 2 sets of global trends are sufficiently close to warrant additional study in search of common fundamental global tectonic controls.

Phanerozoic reconstructions.—The history of continental drift during the past 600 Ma (Phanerozoic) is represented in 13 continental reconstructions at about 50 Ma spacings (fig. 5A, B) after Scotese (1979), Scotese and others (1979), and, in the case of the 600 Ma frame (fig. 5B), Morel and Irving (1978). In each frame, where possible, the three ring-shields and their common tie-line are shown. It is to be stressed that all continental reconstructions and their paleomagnetic basis are equivocal to some extent, and the older the time frame the more equivocal the reconstruction. Thus published younger Phanerozoic reconstructions generally agree, whereas older Phanerozoic reconstructions may vary considerably by author. Despite these reservations, the selected patterns (fig. 5A, B) are considered to provide a meaningful assessment of ring-shield alignment and oscillation during Phanerozoic time.

Available evidence points to the Atlantic as the site of repeated Africa-America Wilson (open-close) Cycles (table 4) (Wilson, 1966). Specifically, oceanic crust or equivalent (ophiolite) is dated at: (1) today (Atlantic), (2) 350 to 550 Ma (Iapetus), (3) 700 to 950 Ma (proto-Iapetus 1), and, possibly, (4) 1100 to 1350 Ma (Proto-Iapetus 2). The data are interpreted in terms of $3\frac{1}{2}$ cycles each of about 400 Ma duration during



(B) A continental reconstruction frame at 600 Ma (Late Proterozoic) after Morel and Irving (1978) illustrating sustained triple-shield alignment.

the past 1400 Ma: Current (0-200 Ma), Appalachian (200-600 Ma), Pan-African (600-1000 Ma), and Grenvillian (1000-1400 Ma) (table 4). The Atlantic Ocean is presently 5000 km wide. Iapetus has been estimated at both 1000 km (Roberts and Gale, 1978) and 1000 to 3000 km wide (Williams, 1980). Proto-Iapetus 1 and 2 ocean widths are highly conjectural; Irving and McGlynn (1976) (see also Irving, Emslie, and Veno, 1974) estimated a Proto-Iapetus 2 width of 5000 km at 1150 Ma ago; however, the same (pre-1200 Ma) ocean opening has been more recently re-evaluated to that of only a "small ocean" (Dunlop and others, 1980). Incorporating the more recent "small ocean" estimate of Dunlop and others (1980), the data on the 4 cycles suggest an increasing amplitude of ocean floor separation with decreasing age across the 4 cycles, that is 5000 km, 1000 to 3000 km, and "small ocean" widths respectively. Two still earlier about 400 Ma duration cycles, of disputed oceanic involvement, are grossly defined by the terminal Elsonian (1.4 Ga) and Hudsonian (1.8 Ga) orogenies (cycles V and VI respectively). These 6 cycles correspond to the long cycles in the Earth's orogenic history earlier described by Vening Meinesz and others as summarized by Gastil (1960).

In summary the data suggest a cyclic pattern of about 400 Ma duration, each cycle about equally divided between (A) early rift-induced extension typically with opening of intervening (Africa-America) ocean basins marking supercontinent break-up and dispersion, and (B) gradual supercontinent reassembly with variable collisional events including major terminal collision (orogeny). According to the paleomagnetic-based Phanerozoic reconstructions (fig. 5A, B), the triple Precambrian ring-shield alignment was maintained during Wilson Cycles II and I (to date), that is, during the past 600 Ma. Additionally during this interval the global continental fragments rotated counter-clockwise (view fig. 5A counter clockwise from the upper left). The positions of the three ring-shields during the pre-600 Ma cycles (that is, III-IV) are highly conjectural and, in effect, unresolved. Piper (1974, 1976) interprets a Mid-Proterozoic (1800-1000 Ma) supercontinent similar to the later short-lived (400-150 Ma) supercontinent Pangea, the one significant variation being that in his Mid-Proterozoic supercontinent, Africa and the rest of Gondwanaland are placed to the "southwest" of North America. However, a simple about 50° sinistral, that is longitudinal (lacking paleomagnetic control), adjustment of Gondwanaland along Pipers' "Hercynian Lineament" (Ouachitas-Anti-Atlas), would restore a Mid-Proterozoic supercontinent closely resembling the later (400-150 Ma) conventional Pangea but with the designated three ring-shield great circle alignment intact. This is, in fact, the preferred Mid-Proterozoic supercontinent reconstruction of Morris, Schmidt, and Roy (1979). In summary available data permit without proving the presence of a single Mid-Proterozoic Pangea-style supercontinent (cycles IV-V).

Outstanding crustal problems, then, revolve about sustained Precambrian ring-shield alignment during repeated crustal oscillations.

Shield roots.—A plausible explanation for maintenance of shield alignment during crustal oscillations emphasizes the role of deep sub-shield mantle roots. Thus seismic evidence has been used to interpret the presence of such roots beneath stable shields and platforms to depths of at least 400 km (Jordan, 1977, 1978, 1979) and of related mantle heterogeneities to depths of 650 to 700 km (Ram and Yadav, 1980; Romanowicz, 1979). Travel-time anomalies are reported to exist in downgoing slabs to depths of 900 to 1000 km (Jordan as reported in *Geotimes*, March 1983, p. 14). The focal mechanisms of earthquakes and the geometry of sinking slabs suggest that there is a barrier to convection at a depth of 700 km (McKenzie and Weiss, 1980). O'Nions, Evensen, and Hamilton (1979) suggest that only the upper mantle has been the source region of the continents, and that the lower mantle was not involved. Thus upper mantle thermal convections and sub-shield roots may both occupy this + 700 km-thick upper mantle-crust interval. According to Jordan (1977) the outer region of the Earth occupied by coherent plates is termed the tectosphere. The thickness of the continental tectosphere is highly variable: it is thin (> 100 km) in unstable continental regions and thick (400 to + 700 km) beneath the stable shields and platforms. Accordingly the deeper sub-shield tectosphere includes substantial mantle thicknesses, integral parts of coherent moving plates. Furthermore, since all thermal models show that the mean temperature of the mantle has decreased with time (McKenzie and Weiss, 1980), it follows that the difference in thickness of continental and oceanic tectosphere was considerably greater in the Precambrian than it is now.

Jordan (1978, 1979) has proposed that continental root zones are generated by removal of basaltic magma from the mantle. This basalt depletion produces a residue less dense than the parental material. Once depleted, low density residual mantle cannot be easily reunited with undepleted mantle. The resulting depleted sub-shield roots are less dense and cooler than adjoining sub-oceanic mantle material. Such a root zone is stabilized at lower temperature, its solidus is elevated, and its effective viscosity is thereby greater. These properties serve to isolate the root zones from many thermal perturbations and mass motions. Furthermore the lower temperatures of the root zones result in horizontal thermal gradients which also contribute to root zone integrity. A thick cool comparatively viscous depleted tectospheric root capped by a mature Precambrian shield-bearing continental platform is envisaged to be the natural product of recurrent magmatic-dynamic processes of the past + 3.8 Ga.

It follows that the development of deep sub-shield roots requires repeated orogenic cycles such as characterize the growth and development of Precambrian shields. The two global shield alignments delineated above, especially the ring-shield-bearing North American-African sector, may be likened to a necklace, each component ring-shield on the "string" having deep coherent roots which, by influencing mantle convections, would serve to preserve both shield integrity and shield alignment on the "string." Sub-shield tectospheric roots were developed and consolidated

episodically, particularly during early Precambrian granitoid-greenstone orogenic cycles (table 1). Sub-shield roots, then, afford a tantalizing plausible link between the common basalt-depleting, root-growing, granitoid-greenstone-producing, shield-stabilizing early Precambrian (> 1.8 Ga) orogenies on the one hand and modern mantle-induced ocean-floor spreading on the other. Root depth (about 700 km), in turn, may have controlled maximum tectonically stable surface shield area (2500-3000 km diam). In summary, ring-shield alignment during both early constructional and later fragmentation stages is tentatively attributed to the presence of deep sub-shield tectospheric roots. Surface wave tomography (Anderson, 1984) offers a critical test of this model.

Oscillations and cycles.—Busse (1978) has proposed an instability mechanism that may be related to episodic (that is, about 400 Ma) continental breakup. Outer Earth comprises an adiabatically stratified viscous fluid, the mantle, covered by a lighter, more viscous material, the crust. The crust is comparatively rich in radioactive heat sources and possesses a lower heat conductivity than the underlying mantle. Under these conditions, an instability of static state can occur as follows: (1) a perturbation causes the crust to become thicker in some places than in others; (2) because of higher heat source and lower heat conductance, the underlying mantle becomes hotter where the crust is thicker; (3) this results in an upward motion where the crust is thicker and downward motion where it is thinner; (4) the stress exerted by motions underneath the crust tends to move crustal materials toward places of downward motion; (5) the crust thickens at these places, and as a result heat production increases above the initially cooler places and decreases above the initially hotter places; and (6) as temperature waves propagate into the underlying mantle the motion reverses itself, and the process is repeated with the opposite sign of motion. The process assumes the character of an instability if the parameters of the problem are such that the amplitude increases in each cycle. Such an oscillatory mechanism provides a plausible explanation for the apparent time dependent nature of plate motions which are the surface manifestation of gravitational instability induced by deep heat production.

As discussed above, continental fragments in the Atlantic region have oscillated, possibly with increasing amplitude, during at least the past 600 Ma ($1\frac{1}{2}$ cycles). The oscillations have apparently operated within the constraints of sustained shield alignment, the dispersed fragments returning to their approximate point of departure. A thermally-induced instability of the type proposed by Busse is a plausible explanation for this 400 Ma-long oscillation. The precision of the reassemblage would vary with thermal convection patterns. Increasing amplitude could conceivably lead to irretrievable dispersal of some continental fragments, for example, Indian Shield. A spectrum of Pangea-type supercontinent reconstructions would result.

Gastil (1960) previously established that intervals of abundant global igneous and metamorphic mineral dates, corresponding to periods of

orogeny, are about 210 Ma in length. These alternate with like intervals of mineral date scarcity (tectonic quiescence) for a mean 417 Ma-long cyclic distribution pattern extending back to 2600 Ma. This pattern corresponds to long cycles in Earth's orogenic history, the peaks in the number of radiometric age corresponding to terminal events of the major crustal processes. Recent dating tends to confirm the validity of Gastil's main peaks of mineral dates marking global orogenies and accelerated crustal growth, the latter recently (Moorbath, 1978; Taylor and others, 1984) augmented to include intervals at 2.9 to 2.6, 3.3 to 3.0, and 3.7 to 3.5 Ga. Culminations of the same crustal processes are also reflected in the post-Archean (2.6 Ga) paleomagnetic record by "hairpin" turns which mark sudden reversals in the sense of polar motion of continental plates (York and Farquhar, 1972). This provides a collective plate tectonic cyclicity which is in accord with Busse's proposed instability mechanism.

Worsley, Moody, and Nance (in press) summarize an attractive non-random crustal model to account for long-term tectonic cyclicity. Plate motion is attributed to the influence on terrestrial heat flow of the repeated assembly of supercontinents (Pangea) that never completely disperse. Briefly, the strong hemispherical asymmetry in the distribution of the ocean (60 percent of Earth's surface) and supercontinent (Pangea) during periods of continental assembly will, due to higher mantle heat flow through oceanic crust and less efficient heat transfer through stationary Pangea, cause an upward displacement of Pangean geothermal gradient that will initially result in uplift of the supercontinent and, ultimately, in hot-spot-induced rifting followed by dispersal of the continental fragments. However, fragment dispersal will continue only until the outer margins of the aging "Atlantic-type" ocean crust ("interior" oceans) become sufficiently cold and dense to "self-subduct" spontaneously, an interval of about 200 Ma. Growth of passive-margined "interior" oceans then is self-limiting and will be overcome presumably by the greater rate of activity of the larger compensating still active-margined parent ocean (Pacific or predecessor). With the onset of subduction in the "interior" oceans, these oceans will again close thereby reassembling another Pangean supercontinent. A pattern of plate tectonic cycles, each cycle of about 400 Ma duration, is recognized as far back as 2000 Ma (Worsley, Nance, and Moody, 1984).

Similarly, LePichon and Huchon (1984) interpret evidence pertaining to the geoid and supercontinent in terms of a weak coupling of a separate steady-state lower mantle which is responsible for the present geoid, to upper mantle convection. This weak coupling leads to hemispheric continental configuration which ends when excessive heating of the upper mantle due to the insulating continental cap leads to continental dispersal. The complete cycle from one supercontinent to the next is considered to be in the order of 400 Ma, a duration coincident with that of the fundamental tectonic oscillations as developed in the present paper.

The role of deeply rooted Precambrian shields in this pattern of non-random plate oscillation is inferred to be in the direction of inhibiting plate dispersal by processes of tectospheric drag, while maintaining original crustal alignments during dispersal and reassemblage.

The contrary, popular "random" plate motion model, however, advocates that Mesozoic Pangea represents assemblage of continental fragments dispersed from a still-earlier supercontinent centered on the Pacific Ocean and now marked by the central Pacific residual geoid high (Anderson, 1982).

Either model illustrates Busse's proposed instability mechanism. Rigorous testing by precise, highly coordinated paleomagnetic-geochronologic techniques is required to investigate continental reconstructions during Precambrian time.

In conclusion Precambrian shield growth, alignment, and oscillation may have operated in accord with repeated or subtly changing thermally-induced instabilities. Precambrian shields may thereby provide insight to mantle convections past and present.

ACKNOWLEDGMENTS

The concept of Canadian-African shield alignments and their relation to tectospheric roots was developed during study leave at Centre National de Recherche Scientifique (CNRS), Montpellier, including field studies in West Africa with the Montpellier group under the leadership of R. Black. I am grateful to the earth scientists of that Institution for their cooperation.

I acknowledge the value of critical reviews of an earlier manuscript by Alfred Kröner and Kevin Burke. Valuable discussion of the physical aspects of basin development and mantle convections with G. T. Jarvis and W. R. Peltier, University of Toronto, are gratefully acknowledged.

An abbreviated version of this paper was presented at the Annual Meeting of the Geological Society of America in 1980. This work has been supported by individual research grants from the National Science and Engineering Research Council of Canada.

REFERENCES

- Affaton, P., Sougy, J., and Trompette, R., 1980, The tectono-stratigraphic relationship between the upper Precambrian and Lower Volta Basin and the Pan-African Dahomeyide orogenic belt (West Africa): *Am. Jour. Sci.*, v. 280, p. 224-248.
- Anderson, D. L., 1984, Surface wave tomography: EOS, *Trans. Am. Geophys. Union*, v. 65, p. 147-148.
- 1982, Hotspots, polar wander, Mesozoic convection and the geoid: *Nature*, v. 297, p. 391-393.
- Anderson, J. L., 1983, Proterozoic anorogenic granite plutonism of North America, in Medaris, L. G., Mickelson, D. M., Byers, C. W., and Shanks, W. C., eds., *Proterozoic Geology: Selected Papers from an International Proterozoic Symposium: Geol. Soc. America Mem.* 161, p. 133-154.
- Baer, A. J., 1976, The Grenville Province in Helikian times: a possible model of evolution: *Royal Soc. London Philos. Trans.*, v. A280, p. 499-515.
- 1981, A Grenvillian model of Proterozoic plate tectonics, in Kröner, A., ed., *Precambrian Plate Tectonics: Amsterdam, Elsevier Sci. Pub. Co.*, p. 353-385.
- Bally, A. N., 1980, Basin and subsidence — a summary, in Bally, A. W., Bender, P. L., McGetchin, T. R., and Walcott, R. I., eds., *Dynamics of Plate Interiors (Geodynamics Ser., v. 1): Washington, D.C., Am. Geophys. Union/Boulder, Colo., Geol. Soc. America*, p. 5-20.

- Baragar, W. R. A., and Scoates, R. F. J., 1981, The circum-Superior belts: a Proterozoic plate margin?, *in* Kroner, A., ed., *Precambrian Plate Tectonics*: Amsterdam, Elsevier Sci. Pub. Co., p. 297-330.
- Bell, R. T., 1970, The Hurwitz Group, a prototype for deposition on metastable cratons, *in* Baer, A. J., ed., *Basins and Geosynclines of the Canadian Shield*: Canadian Geol. Survey Paper 70-40, p. 159-169.
- Bertrand, J. M. L., and Caby, R., 1978, Geodynamic evolution of the Pan-African orogenic belt: a new interpretation of the Hoggar Shield (Algerian Sahara): *Geol. Rundschau*, v. 67, no. 2, p. 357-388.
- Bertrand, J. M. L., and Lasserre, M., 1976, Pan-African and pre-Pan-African history of the Hoggar (Algerian Sahara) in the light of new geochronological data from the Aleksod area: *Precambrian Research*, v. 3, p. 343-362.
- Bertrand-Sarfati, J., 1972, Stromatolites columnaires de certaines formations carbonatées du Précambrien supérieur du bassin congolais: Tervuren, Belgium, Mus. Royal de l'Afrique Centrale, *Annales Ser 8, Sci. Géol.* no. 74, p. 1-45.
- Bertrand-Sarfati, J., Fabre, J., and Moussine-Pouchkine, A., 1977, *Geodynamique des aires sedimentaires cratonique; quelques exemples sahariens*: Bull. Centres Recherche Explor. -Prod. Elf-Aquitaine, v. 1, p. 217-231.
- Bessoles, B., 1977, *Geologie de l'Afrique: le craton ouest africain*: Mem. Bur. Recherches Geol. et Minières, no. 88, 402 p.
- Bessoles, B., and Trompette, R., 1980, *Geologie de l'Afrique. La Chaîne panafricaine "zone mobile d'Afrique centrale (partie sud) et zone mobile soudanaise"*: Mem. Bur. Recherches Geol. et Minières, no. 92, 397 p.
- Black, R., 1980, *Precambrian of West Africa: Episodes*, no. 4, p. 3-8.
- Blake, D. H., 1980, *Volcanic rocks of the Paleohelikian Dubawnt Group in the Baker Lake — Angikuni Lake area, District of Keewatin, Northwest Territories*: Canada Geol. Survey Bull. 309, 39 p.
- Bonhomme, M. G., and Bertrand-Sarfati, J., 1982, *Correlation of Proterozoic sediments of western and central Africa and South America based upon radiochronological and palaeontological data*: *Precambrian Research*, v. 18, p. 171-194.
- Bonhomme, M. G., Gauthier-Lafage, F., and Weber, F., 1982, *An example of Lower Proterozoic sediments: the Francevillian in Gabon*: *Precambrian Research*, v. 18, p. 87-102.
- Bonner, G., Roussel, J., and Trompette, R., 1980, *Genesis and geodynamic evolution of The Taoudeni cratonic basin, (Upper Precambrian and Paleozoic), Western Africa*, *in* Bally, A. W., Bender, P. L., McGetchin, T. R., and Walcott, R. I., eds., *Dynamics of Plate Interiors (Geodynamics Ser., v. 1)*: Am. Geophys. Union/Geol. Soc. America, p. 81-90.
- Bullard, E., Everett, J. E., and Smith, A. G., 1965, *The fit of continents around the Atlantic*: *Royal Soc. London Philos. Trans., Ser. A* 258, p. 41-51.
- Burke, K. C., and Wilson, J. T., 1976, *Hot-spots on the Earth's surface*: *Sci. Am.*, v. 235, p. 46-57.
- Burke, K. C., and Dewey, J. F., 1973, *Plume-generated triple junctions: Key indicators in applying plate-tectonics to old rocks*: *Jour. Geology*, v. 81, p. 406-433.
- Busse, F. H., 1978, *A model of time-periodic mantle flow*: *Royal Astron. Soc., Jour. Geophysics*, v. 53, p. 1-12.
- Caby, R., Bertrand, J. M., and Black, R., 1981, *Pan-African ocean closure and collision in the Hoggar — Iforas segment, central Sahara*, *in* Kroner, A., ed., *Precambrian Plate tectonics*: Holland, Elsevier, p. 407-434.
- Cahen, L., 1982, *Geochronological correlation of the Late Precambrian sequences on and around the stable zones of Equatorial Africa*: *Precambrian Research*, v. 18, p. 73-86.
- Cahen, L., Snelling, N. J., Delhal, J., and Vail, J. R., 1984, *The geochronology and evolution of Africa*: Oxford, Clarendon Press, 512 p.
- Cahen, L., Delhal, J., and Lavreau, J., 1976, *The Archean of Equatorial Africa: a review*, *in* Windley, B. F., ed., *The Early History of the Earth*: London, Wiley, p. 489-498.
- Cahen, L., and Lepersonne, J., 1967, *The Precambrian of the Congo, Rwanda, and Burundi*, *in* Rankama, K., ed., *The Precambrian 3*: New York, Intersci. Pub., p. 143-290.
- Cahen, L., and Snelling, N. J., 1966, *The geochronology of Equatorial Africa*: Amsterdam, North Holland Publishing Co., 195 p.
- Cambray, F. W., 1978, *Plate tectonics as a model for the environment of deposition and deformation of the early Proterozoic (Precambrian X) of northern Michigan [abs.]*: *Geol. Soc. America Abs. with Programs*, v. 7, no. 7, p. 376.

- Campbell, F. H. A., and Cecile, M. P., 1981, Evolution of the early Proterozoic Kilohigok Basin, Bathurst Inlet-Victoria Island, Northwest Territories, *in* Campbell, F. H. A., ed., Proterozoic Absins of Canada: Canada Geol. Survey, Paper 81-10, p. 103-131.
- Capdevila, R., Goodwin, A. M., Ujike, S., and Gorton, M. J., 1982, Trace-element geochemistry of Archean volcanic rocks and crustal growth in southwestern Abitibi Belt, Canada: *Geology*, v. 10, p. 418-422.
- Card, K. D., Church, W. R., Franklin, J. M., Frarey, M. J., Robertson, J. A., West, G. F., and Young, G. M., 1972, The Southern Province, *in* Price, R. A., and Douglas, R. J. W., eds., Variations in Tectonic Style: Geol. Assoc. Canada Spec. Paper 11, p. 335-380.
- Chandler, F. W., 1982, The structure of the Richmond Gulf Graben and the geological environments of lead-zinc mineralization and of iron-manganese formation in the Nastapoka Group, Richmond Gulf area, New Quebec-Northwest Territories, *in* Current Research, pt. A: Canada Geol. Survey Paper 82-1A, p. 1-10.
- Chikhaoui, M., Dupuy, C., and Dostal, J., 1980, Geochemistry and petrogenesis of Late Proterozoic volcanic rocks from north-western Africa: *Contr. Mineralogy Petrology*, v. 73, p. 375-388.
- Choubert, G., and Faure-Muret, A., 1971, Tectonique de l'Afrique: UNESCO, *Sci. de la Terre* 6, 602 p.
- Clauer, N., 1976, Geochimie isotopique du strontium des milieux sedimentaire: Univ. Louis Pasteur de Strasbourg, *Inst. Geologie, Mem.* 45, 256 p.
- Clauer, R., Caby, R., Daniel, J., and Trompette, R., 1982, Geochronology of sedimentary and metasedimentary Precambrian rocks of the West African Craton: *Precambrian Research*, v. 18, p. 53-71.
- Clifford, T. N., 1968, Radiometric dating and pre-Silurian geology of Africa, *in* Hamilton, E. I., and Farquhar, R. M., eds., Radiometric dating for geologists: London, Intersci. Pub., p. 299-416.
- 1970, The structural framework of Africa, *in* Clifford, T. N., and Gass, I., eds., African Magmatism and Tectonics: Edinburgh, Oliver and Boyd, p. 1-26.
- 1974, Review of African granulites and related rocks: *Geol. Soc. America Spec. Paper* 156, 49 p.
- Collerson, K. D., Jesscau, C. W., and Bridgewater, D., 1975, Crustal development of the Archean gneiss complex: eastern Labrador, *in* Windley, B. F., ed., *The Early History of the Earth*: London, John Wiley & Sons, p. 237-253.
- Davidson, A., 1972, The Churchill Province, *in* Price, R. A., and Douglas, R. J. W., eds., Variations in Tectonic Style in Canada: Geol. Assoc. Canada Spec. Paper 11, p. 381-433.
- Davidson, A., Culshaw, N. G., and Nadeau, L., 1982, A tectono-metamorphic framework for part of the Grenville Province, Parry Sound region, Ontario, *in* Current Research, pt. A: Canada Geol. Survey Paper 82-1A, p. 175-190.
- Davis, D. W., and Sutcliffe, R. H., 1984, U-Pb ages from the Nipigon Plate: *Geol. Assoc. Canada Abs. with Program*, v. 9, p. 57.
- Dearnley, R., 1966, Orogenic fold belts and a hypothesis of Earth evolution, *in* Ahrens, L. H., Press, F., Runcorn, J. K., and Vreg, H. C., eds., *Physics and Chemistry of the Earth*: London, Pergamon Press, p.
- Dewey, J. F., and Burke, K. C. A., 1974, Hot spots and continental break-up: implications for collisional orogeny: *Geology*, v. 2, p. 57-60.
- Dietz, R. S., and Holden, J. C., 1970, Reconstruction of Pangaea: break-up and dispersion of continents, Permian to Present: *Jour. Geophys. Research*, v. 75, p. 4939-4956.
- Dimroth, E., 1970, Evolution of the Labrador geosyncline: *Geol. Soc. America Bull.*, v. 81, p. 2717-2741.
- 1981, Labrador Geosyncline: type example of early Proterozoic cratonic reactivation, *in* Kröner, A., ed., *Precambrian Plate Tectonics*: Amsterdam, Elsevier, p. 331-352.
- 1982, Evolution of the south-central part of the Archean Abitibi Belt, Quebec. Part I: stratigraphy and paleogeographic model: *Canadian Jour. Earth Sci.*, v. 19, p. 1729-1758.
- Dodson, M. H., Bell, K., and Shackleton, R. M., 1973, Archean geochronology of East Africa: *Fortschr. Mineralogie*, v. 50, p. 67-68.
- Donnet, D., 1976, Some aspects of the Panarctic cratonic margin in Australia: *Royal Soc. London Philos. Trans.*, v. A280, p. 641-654.
- Douglas, R. J. W., 1970, Geology and Economic Minerals of Canada: Canada Geol. Survey, *Econ. Geology Rept.* 1, 838 p.

- Dunlop, D. J., York, D., Berer, G. W., Buchan, K. L., and Stirling, J. M., 1980, The Grenville Province: a paleomagnetic case-study of Precambrian continental drift, *in* Strangway, D. W., ed., *The Continental crust and its Mineral Deposits*: Geol. Assoc. Canada Spec. Paper 20, p. 487-502.
- Engel, A. E. J., and Kelm, D. L., 1972, Precambrian global tectonics: a tectonic test: *Geol. Soc. America Bull.*, v. 83, p. 2325-2340.
- Fahrig, W. F., Christie, K. W., Chowin, E. H., James, D., and Machado, N., 1984, Tectonic significance of some major Aphebian diabase dyke swarms: *Geol. Assoc. Canada Abs. with Program*, v. 9, p. 61.
- Fahrig, W. F., Christie, K. W., and Jones, D. L., 1981, Paleomagnetism of the Bylot Basin: evidence for Mackenzie continental tensional tectonics, *in* Campbell, F. H. A., ed., *Proterozoic Basins of Canada*: Canada Geol. Survey Paper 81-10, p. 303-312.
- Fraser, J. A., 1978, Metamorphism in the Churchill Province, District of Mackenzie, *in* Fraser, J. A., and Heywood, W. W., eds., *Metamorphism in the Canadian Shield*: Canada Geol. Survey Paper 78-10, p. 195-202.
- Fraser, J. A., Donaldson, J. A., Fahrig, W. F., and Tremblay, L. P., 1970, Helikian basins and geosynclines of the northwestern Canadian Shield, *in* Baer, A. J., ed., *Symposium on Basins and Geosynclines in the Canadian Shield*: Canada Geol. Survey Paper 70-40, p. 213-238.
- Fraser, J. A., and Heywood, W. W., 1978, *Metamorphism in the Canadian Shield*: Canada Geol. Survey Paper 78-10, 367 p.
- Fraser, J. A., Hoffman, P. F., Irvine, T. N., and Mursky, G., 1972, The Bear Province, *in* Price, R. A., and Douglas, R. J. W., eds., *Variations in Tectonic Style*: Geol. Assoc. Canada Spec. Paper 11, p. 453-503.
- Frisch, T., 1982, Precambrian geology of the Prince Albert Hills, western Melville Peninsula, Northwest Territories: *Canada Geol. Survey Bull.* 346, 70 p.
- Gastil, G., 1960, The distribution of mineral dates in time and space: *Am. Jour. Sci.*, v. 258, p. 1-35.
- Gibb, R. A., and Walcott, R. I., 1971, A Precambrian suture in the Canadian Shield: *Earth Planetary Sci. Letters*, v. 10, p. 417-422.
- Goldich, S. S., Hedge, C. E., and Stern, 1970, Age of the Morton and Montevideo gneisses and related rocks, southwestern Minnesota: *Geol. Soc. America Bull.*, v. 81, p. 3671-3696.
- Goodwin, A. M., 1973, Plate tectonics and evolution of Precambrian crust, *in* Tarling, D. H., and Runcorn, S. K., eds., *Implications of Continental Drift to the Earth Sciences*: London, Academic Press, p. 1047-1069.
- 1974, Precambrian belts, plumes and shield development: *Am. Jour. Sci.*, v. 274, p. 987-1028.
- 1978, The nature of Archean crust in the Canadian Shield, *in* Tarling, D. H., ed., *Evolution of the Earth's Crust*: London, Academic Press, p. 175-218.
- 1980, Precambrian ring-shields and the Atlantic penning [abs.]: *Geol. Soc. America Ann. Mtg.*, v. 12, p. 435.
- 1981a, Archean plates and greenstone belts, *in* Kroner, A., ed., *Precambrian Plate Tectonics*: Amsterdam, Elsevier, p. 105-135.
- 1981b, Precambrian perspectives: *Science*, v. 213, p. 55-61.
- Goodwin, A. M., Ambrose, J. W., Ayres, L. D., Clifford, P. M., Currie, K. L., Ermanovics, I. M., Fahrig, W. F., Gibb, R. A., Hall, D. H., Innes, M. J. S., Irvine, T. N., MacLaren, A. S., Norris, A. W., Pettijohn, F. J., and Ridler, R. H., 1972, The Superior Province, *in* Price, R. A., and Douglas, R. J. W., eds., *Variations in Tectonic Styles in Canada*: Geol. Assoc. Canada Spec. Paper 11, p. 527-623.
- Grant, N. K., 1978, Structural distinction between a metasedimentary cover and an underlying basement in the 600 m.y. old Pan-African domain of northwestern Nigeria, West Africa: *Geol. Soc. America Bull.*, v. 89, p. 50-58.
- Hedge, C. E., Marvin, R. F., and Naeser, C. W., 1975, Age provinces in the basement rocks of Liberia: *U.S. Geol. Survey Jour. Research*, v. 3, p. 425-429.
- Heaman, L. M., Shaw, D. M., and McNutt, R. H., 1982, The Tallan Lake Sill: geochemical evidence for large-scale nappe structure in the NW portion of the Grenville Province [abs.]: *Geol. Soc. America*, v. 14, no. 7, p. 512.
- Henderson, J. B., 1981, Archean basin evolution in the Slave Province, Canada, *in* Kroner, A., ed., *Precambrian Plate Tectonics*: Amsterdam, Elsevier, p. 213-235.
- Hepworth, J. V., 1967, Photogeological recognition of ancient orogenic belts in Africa: *Geol. Soc. London Quart. Jour.*, v. 123, p. 253-292.
- Hertz, N., 1969, Anorthosite belts, continental drift and the anorthosite event: *Science*, v. 164, p. 944-947.

- Heywood, W. W., and Schau, M., 1981, Geology of Baker Lake Region, District of Keewatin: Canada Geol. Survey Paper 81-1A, p. 259-264.
- Hoffman, P. F., 1980, Wopmay Orogen: A Wilson Cycle of Early Proterozoic age in the northwest of the Canadian Shield, in Strangway, W. W., ed., *The Continental Crust and its Mineral Deposits*: Geol. Assoc. Canada Spec. Paper 20, p. 523-549.
- Hoffman, P. F., and Pelletier, K. S., 1982, Cloos Nappe in Wopmay Orogen: significance for stratigraphy and structure of the Akaitcho Group, and implications for opening and closing of an early Proterozoic continental margin, in *Current Research*, part A: Canada Geol. Survey Paper 82-1A, p. 109-115.
- Hoffman, P. F., Card, K. D., and Davidson, A., 1982, The Precambrian: Canada and Greenland: Geol. Soc. America, D-NAG Spec. Pub. 1, p. 3-6.
- Hoffman, P. F., and St. Onge, M. R., 1981, Contemporaneous thrusting and conjugate transect faulting during the second collision in Wopmay Orogen: implications for the subsurface structure of post-orogenic outliers, in *Current Research*, Pt. A: Canada Geol. Survey, Paper 81-1A, p. 251-257.
- Holst, T. B., 1984, Evidence for nappe development during the early Proterozoic Penokean orogeny, Minnesota: *Geology*, v. 12, p. 13-138.
- Hurley, P. M., 1970, Distribution of age provinces in Laurasia: *Earth Planetary Sci. Letters*, v. 8, p. 189-196.
- 1974, Pangeaic orogenic system: *Geology*, v. 2, p. 373-376.
- Hurley, P. M., Fairbairn, H. W., and Gaudette, H. E., 1976, Progress report on early Archean rocks in Liberia, Sierra Leone and Guayana, and their general stratigraphic setting, in Windley, B. F., ed., *The Early History of the Earth*: London, John Wiley & Sons, p. 511-521.
- Irving, E., Emslie, R. F., and Ueno, H., 1974, Upper Proterozoic paleomagnetic poles from Laurentia and the history of the Grenville Structural Province: *Jour. Geophys. Research*, v. 79, p. 5491-5502.
- Irving, E., and McGlynn, J. C., 1976, Proterozoic magnetostratigraphy and the tectonic evolution of Laurentia: *Royal Soc. London Philos. Trans.*, ser. A, v. 280, p. 433-468.
- Jackson, G. D., and Iannelli, T. R., 1981, Rift-related cyclic sedimentation in the Neohelikian Borden Basin, Northern Baffin Island, in Campbell, F. H. A., ed., *Proterozoic Basins of Canada*: Canada Geol. Survey Paper 81-10, p. 269-302.
- Jackson, G. D., and Morgan, W. C., 1978, Precambrian metamorphism in Baffin and Bylot Islands, in Fraser, J. A., and Heywood, W. W., eds., *Metamorphism in the Canadian Shield*: Canada Geol. Survey Paper 78-10, p. 249-267.
- Jordan, T. H., 1977, Lithospheric slab penetration into the lower mantle beneath the Sea of Okhotsk: *Jour. Geophysics*, v. 43, p. 473-496.
- 1978, Composition and development of the continental tectosphere: *Nature*, v. 274, p. 544-548.
- 1979, The deep structure of the continents: *Sci. Am.*, v. 240, p. 70-82.
- Kennedy, W. Q., 1964, The structural differentiation of Africa in the Pan-African (500 Ma) tectonic episode: Univ. Leeds, Research Inst. African Geology, 8th Ann. Rept., p. 48-49.
- Kerr, J. W., 1980, A plate tectonic contest in Arctic Canada, in Strangway, D. W., ed., *The Continental Crust and its Mineral Deposits*: Geol. Assoc. Canada Spec. Paper 20, p. 457-486.
- Kerans, C., Ross, G. M., Donaldson, J. A., and Geldsetzer, H. J., 1981, Tectonism and depositional history of the Helikian Hornby Bay and Dismal Lakes groups, District of Mackenzie, in Campbell, F. H. A., ed., *Proterozoic Basin of Canada*: Canada Geol. Survey Paper 81-10, p. 157-182.
- Klasner, J. S., Cannon, W. F., and Van Schmus, W. R., 1982, The pre-Keweenaw tectonic history of southern Canadian shield and its influence on formation of the Midcontinent Rift: Geol. Soc. America Mem. 156, p. 27-46.
- Klasner, J. S., and Schultz, K. J., 1982, Concentrically zoned pattern in the Bouguer gravity anomaly map of northeastern North America: *Geology*, v. 10, p. 537-541.
- Krough, T. E., David, D. W., and Corfu, F., 1984, Implications of precise U-Pb dating for the geological evolution of the Superior Province: Geol. Assoc. Canada Abs. with Program, v. 9, p. 79.
- Krough, T. E., McNutt, R. H., and Davis, G. L., 1981, Two high precision U-Pb zircon ages for the Sudbury Nickel Irruptive: *Canadian Jour. Earth Sci.*, v. 19, p. 723-728.

- Kröner, A., 1977a, Precambrian mobile belts of southern and eastern Africa — ancient sutures or sites of ensialic mobility? A case for crustal evolution towards plate tectonics: *Tectonophysics*, v. 40, p. 101-135.
- 1977b, The Precambrian geotectonic evolution of Africa: plate accretion versus plate destruction: *Precambrian Research*, v. 4, p. 163-213.
- 1980, Pan-African crustal evolution: *Episodes*, no. 2, p. 3-8.
- 1981, Precambrian plate tectonics, in Kröner, A., ed., *Precambrian Plate Tectonics*: Amsterdam, Elsevier, p. 57-90.
- Leblanc, M., 1976, Proterozoic oceanic crust at Bon Azzer: *Nature*, v. 261, p. 34-35.
- 1981, The Late Proterozoic ophiolites of Bou Azzer (Morocco): evidence for Pan-African plate tectonics, in Kronre, A., ed., *Precambrian Plate Tectonics*: Amsterdam, Elsevier, p. 435-451.
- Lepersonne, J., 1977, Structure géologique du bassin intérieur du Zaïre: *Acad. Royale Belgique, Bull. de Classe de Sci., 5^e ser.* — Tome LXIII, p. 941-965.
- LePichon, X., and Huchon, P., 1984, Geoid, Pongea and convection: *Earth Planetary Sci. Letters*, v. 67, p. 123-135.
- Lewry, J. F., and Sibbald, T. I. I., 1980, Thermotectonic evolution of the Churchill Province in northern Saskatchewan: *Tectonophysics*, v. 68, p. 45-82.
- Lewry, J. F., Stauffer, M. R., and Fumerton, S., 1981, A Cordillera-type batholithic belt in the Churchill Province in Northern Saskatchewan: *Precambrian Research*, v. 14, p. 277-313.
- McCulloch, M. T., and Wasserburg, G. J., 1978, Sm-Nd and Rb-Sr chronology of continental crust formation: *Science*, v. 200, p. 1003-1011.
- McElhinny, M. W., Taylor, S. R., and Stevenson, D. J., 1978, Limits to the expansion of Earth, Moon, Mars and Mercury and to changes in the gravitational constant: *London, Nature*, v. 271, p. 316-321.
- McGlynn, J. C., and Henderson, J. B., 1972, The Slave Province, in Price, R. A., and Douglas, R. J. W., eds., *Variations in Tectonic Style in Canada*: *Geol. Assoc. Canada Spec. Paper 11*, p. 505-526.
- McGlynn, J. C., and Irving, E., 1981, Horizontal motions and rotations in the Canadian Shield during the Early Proterozoic, in Campbell, F. H. A., ed., *Proterozoic Basins in Canada*: *Canada Geol. Survey Paper 81-10*, p. 183-190.
- McKenzie, D., and Weiss, N., 1980, The thermal history of the earth, in Strangway, D. W., ed., *The Continental Crust and Its Mineral Deposits*: *Geol. Assoc. Canada Spec. Paper 20*, p. 575-590.
- Mestraud, J. L., 1971, Les couvertures de plate-forme, in *Tectonique de l'Afrique*: Paris, UNESCO, p. 495-507.
- Moorbath, S., 1978, Age and isotope evidence for the evolution of continental crust: *Royal Soc. London Philos. Trans.*, v. 288, p. 401-413.
- Morel, P., and Irving, E., 1978, Tentative paleocontinental maps for the early Phanerozoic and Proterozoic: *Jour. Geology*, v. 86, p. 535-561.
- Morris, W. A., Schmidt, P. W., and Roy, J. L., 1979, A graphical approach to polar paths: paleomagnetic cycles and global tectonics: *Physics of the Earth and Planetary Interiors*, v. 19, p. 85-99.
- Muelberger, W. R., 1980, The shape of North America during the Precambrian, in *Continental Tectonics*: Washington, D.C., *Natl. Acad. Sci.*, p. 175-183.
- Muehlberger, W. R., Denison, R. E., and Lidiak, E. G., 1967, Basement rocks in continental interior of United States: *Am. Assoc. Petroleum Geologist Bull.*, v. 51, p. 2351-2380.
- O'Driscoll, E. S. T., 1980, The double helix in global tectonics: *Tectonophysics*, v. 63, p. 397-417.
- O'Nions, R. K., Evensen, N. M., and Hamilton, P. J., 1979, Geochemical modeling of mantle differentiation and crustal growth: *Jour. Geophys. Research*, v. 84, p. 6091-6101.
- Percival, J. A., and Card, K. D., 1983, Archean crust as revealed in the Kapuskasing uplift, Superior Province Canada: *Geology*, v. 11, p. 323-326.
- Percival, J. A., and Coe, K., 1981, Parallel evolution of Archean low- and high-grade terrane: a view based on relationships between the Abitibi, Wana and Kapuskasing belts: *Precambrian Research*, v. 14, p. 315-331.
- Piper, J. D. A., 1974, Proterozoic crustal distribution, mobile belts and apparent polar movement: *Nature*, v. 251, p. 381-384.
- 1976, Paleomagnetic evidence for a Proterozoic super-continent: *Royal Soc. London Philos. Trans.*, v. A280, p. 469-490.
- Pretorius, D. A., 1973, The crustal architecture of southern Africa: *Geol. Soc. South Africa*, annexure to v. LXXVI, 32 p.

- Price, R. A., and Douglas, R. J. W., 1972, Variations in Tectonic Styles in Canada: Geol. Assoc. Canada Spec. Paper 11, 688 p.
- Raemaker, P., 1981, Hudsonian and Helikian basins of the Athabasca region, northern Saskatchewan, in Campbell, F. H. A., ed., Proterozoic Basins of Canada: Canada Geol. Survey Paper 81-10, p. 219-233.
- Ram, A., and Yadav, L., 1980, Evidence for upper mantle heterogeneity around 650 km depth beneath the Indian subcontinent: Tectonophysics, v. 68, p. T17-T23.
- Rast, N., O'Brien, B. H., and Wardle, R. J., 1976, Relationships between Precambrian and lower Paleozoic rocks of the Avalon Platform in New Brunswick, the northern Appalachians and the British Isles: Tectonophysics, v. 30, p. 315-338.
- Ricketts, B. D., and Donaldson, J. A., 1981, Sedimentary history of the Belcher Group of Hudson Bay, in Campbell, F. H. A., ed., Proterozoic Basins of Canada: Canada Geol. Survey Paper 81-10, p. 235-254.
- Ridler, R. H., 1972, Volcanic stratigraphy and metallogeny of the Kaminak Group, in Report of Activities, pt. A: Canada Geol. Survey Paper 71-1, pt. A, p. 142-148.
- Roberts, David, and Gale, G. H., 1978, The Caledonian-Appalachian Iapetus Ocean, in Tarling, D. H., ed., Evolution of the Earth's Crust: London, Academic Press, p. 255-342.
- Romanowicz, B. A., 1979, Seismic structure of the upper mantle beneath the United States by three-dimensional inversion of body wave arrival times: Royal Astron. Soc. Geophys. Jour., v. 57, p. 479-506.
- Saggerson, E. P., 1978, Metamorphic Map of Africa, with explanatory notes: Paris, CGMW/CCGM, UNESCO, 33 p.
- Schau, M., 1978, Metamorphism of the Prince Albert Group, District of Keewatin, in Fraser, J. A., and Hegwood, W. W., eds., Metamorphism in the Canadian Shield: Canada Geol. Survey Paper 78-10, p. 203-213.
- Scotese, C. R., 1979, Continental Drift Flip Charts, 2d ed.: Univ. Chicago, Dept. Geophys. Sci.
- Scotese, C. R., Bambach, R. K., Barton, C., Vandor Voo, R., and Ziegler, A. M., 1979, Paleozoic base maps: Jour. Geology, v. 87, p. 217-277.
- Sims, P. K., Card, K. D., and Lumbers, S. B., 1981, Evolution of early Proterozoic basins of the Great Lakes region, in Campbell, F. H. A., ed., Proterozoic Basins of Canada: Canada Geol. Survey Paper 81-10, p. 379-397.
- St. Onge, M. R., King, J. E., and Lalonde, A. E., 1982, Geology of the central Wompay Orogen, Bear Province, District of Mackenzie: Redrock Lake and eastern portion of Calder River map area, in Current Research, Pt. A: Canada Geol. Survey Paper 82-1A, p. 99-108.
- Stewart, J. H., 1976, late Precambrian evolution of North America: plate tectonics implication: Geology, v. 4, p. 11-15.
- Stockwell, C. H., 1961, Structural provinces, orogenies and time classification of rocks of the Canadian Precambrian Shield: Canada Geol. Survey Paper 61-17, p. 108-118.
- Strong, D. F., 1979, Proterozoic tectonics of northwestern Gondwanaland: new evidence from eastern Newfoundland: Tectonophysics, v. 54, p. 81-101.
- Tanner, P. W. G., 1973, Orogenic cycles of East Africa: Geol. Soc. America Bull., v. 84, p. 2839-2850.
- Taylor, P. N., Chadwick, B., Moorbath, S., Ramakrishnan, M., and Viswanath, M. N., 1984, Petrology, chemistry and isotopic ages of Peninsular Gneiss, Dharwar acid volcanic rocks and the Chitradorga Granite with special reference to the Late Archean evolution of the Karnataka Craton, Southern India: Precambrian Research, v. 23, p. 349-375.
- Thomas, M. D., and Kearney, P., 1980, Gravity anomalies, block faulting and Archean-type tectonism in the eastern Churchill Province: Nature Phys. Sci., v. 282, p. 61-63.
- Turcotte, D., 1980, Models for the evolution of sedimentary basins, in Bally, A. W., Bender, P. L., McGetchin, T. R., and Walcott, R. I., eds., Dynamics of Plate Interiors (Geodynamics Series, v. 1): Am. Geophys. Union/Geol. Soc. America, p. 21-26.
- Turner, D. C., 1971, The Precambrian and Lower Paleozoic basement of Nigeria, in Tectonics of Africa: UNESCO, Earth Sci., 6, p. 255-260.
- Umbgrove, J. H. F., 1947, The Pulse of the Earth, 2d ed.: The Hague, M. Nijoff, 358 p.
- Van Schmus, W. R., 1976, Early and Middle Proterozoic history of the Great Lakes area, North America: Royal Soc. London Philos. Trans., v. 280, p. 605-628.
- Van Schmus, W. R., and Bickford, M. E., 1981, Proterozoic chronology and evolution of the mid-Continent region: North America, in Kröner, A., ed., Precambrian Plate Tectonics: Amsterdam, Elsevier, p. 261-296.
- Veevers, J. J., and McElhinney, M. W., 1976, The separation of Australia from other continents: Earth-Science Reviews, v. 12, p. 139-159.

- Wallace, H., 1981, Keweenaw geology of the Lake Superior Basin, in Campbell, F. H. A., ed., Proterozoic Basins of Canada: Canada Geol. Survey Paper 81-10, p. 399-417.
- Wardle, R. J., and Bailey, D. G., 1981, Early Proterozoic sequences in Labrador, in Campbell, F. H. A., ed., Proterozoic Basins of Canada: Canada Geol. Survey Paper 81-10, p. 331-358.
- Weber, W., and Scoates, R. F. J., 1978, Archean and Proterozoic metamorphism in the northwestern Superior Province and along the Churchill-Superior boundary, Manitoba, in Fraser, J. A., and Heywood, W. W., eds., Metamorphism in the Canadian Shield: Canada Geol. Survey Paper 78-10, p. 5-16.
- Williams, H. R., 1978a, The Archean geology of Sierra Leone: Precambrian Research, no. 6, p. 251-268.
- 1978b, African Archean mobile belts and granite-greenstone terrain: Nature, v. 266, p. 163-164.
- 1980, Structural telescoping across the Appalachian orogen and the minimum width of the Iapetus Ocean, in Strangway, D. W., ed., The Continental crust and its Mineral Deposits: Geol. Assoc. Canada Spec. Paper 30, p. 421-440.
- Wilson, J. T., 1949, The origin of continents and Precambrian history: Royal Soc. Canada Trans., v. 43, p. 157-182.
- 1966, Did the Atlantic Ocean close then reopen? Nature, v. 211, p. 676-681.
- 1968a, Static or mobile earth, in Gondwana Revisited: Am. Philos. Soc. Proc., v. 112, p. 309-320.
- 1968b, Comparison of Hudson Bay Arc with some other feature, in Beales, C. S., ed., Science History and Hudson Bay: Canada Dept. Energy, Mines and Resources, v. 2, p. 1015-1033.
- Worsley, T. R., Moody, J. B., and Nance, R. D., in press, Proterozoic to Recent tectonic tuning of biochemical cycles, in Sundquist, E. T., ed., Natural Variations in Carbon Dioxide and the Carbon Cycle: Am. Geophys. Union Mon., Carbon Cycle volume.
- Worsley, T. R., Nance, D., and Moody, J. B., 1984, Global tectonics and eustasy for the past 2 billion years: Marine Geology, v. 58, p. 373-400.
- Wynne-Edwards, H. R., 1972, The Grenville Province, in Price, R. A., and Douglas, R. J. W., eds., Variations in Tectonic Styles in Canada: Geol. Assoc. Canada Spec. Paper 11, p. 264-334.
- 1976, Proterozoic ensialic orogenesis: the millipede model of ductile plate tectonics: Am. Jour. Sci., v. 276, p. 927-953.
- York, D., and Farquhar, R. M., 1972, The Earth's Age and Geochemistry: New York, Pergamon Press, 178 p.