Geology of the Venezuelan Guayana Shield and Its Relation to the Geology of the Entire Guayana Shield

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ABSTRACT

The Guayana Shield in Venezuela is composed of five lithotectonic provinces: (1) an Archean amphibolite- to granulite-facies gneiss terrane, (2) an Early Proterozoic greenstone-granite terrane(s), (3) an Early Proterozoic metamorphosed volcanic-plutonic complex, (4) Early to Middle Proterozoic continental sedimentary rocks, and (5) Middle Proterozoic anorogenic rapikivi-type granite. Early Proterozoic rocks in Estado Amazonas of western Venezuela are undivided, and their relation to other rocks of the Venezuelan Guayana Shield is uncertain. Early to Middle Proterozoic continental-type tholeiitic dikes, sills, and small irregular intrusive bodies and Mesozoic dikes emplaced during the opening of the Atlantic Ocean cut all of the lithotectonic provinces. Major mineral deposits of the Venezuelan Guayana Shield include gold, iron, bauxite, and diamonds.

The Archean Imataca Complex, the oldest unit of the shield, consists of gneiss and granulite and minor dolomite and banded iron formation. Large isoclinal folds that have been refolded into relatively open folds are common. Metamorphic grade ranges from granulite facies in the northeast part of the belt to amphibolite facies in the southwest. Deposits of enriched banded iron formation in the Imataca Complex contain more than 2 billion metric tons of iron ore. During the pre-Trans-Amazonian tectonomagmatic event, about 2,800–2,700 Ma, granitic rocks intruded the Imataca Complex, and injection gneiss and migmatite were developed.

The Early Proterozoic greenstone belts, which formed about 2,250–2,100 Ma, consist of a submarine sequence of tholeiitic mafic volcanic rocks, a sequence of tholeiitic to calc-alkaline basalt to rhyolite, and a sequence of turbiditic graywacke, volcaniclastic rocks, and chemical sedimentary rocks that characterize the basal, middle, and upper parts, respectively. Layered mafic complexes also are present in the greenstone belts. Metamorphic grade ranges from greenschist to amphibolite facies. Shear zones that cut the greenstone-belt rocks host numerous deposits of low-sulfide gold-quartz veins.

Granitic domes of the Supamo Complex intruded the greenstone-belt rocks about 2,230–2,050 Ma, dividing the greenstone-belt rocks into branching synclinoria between intrusions. The Trans-Amazonian orogeny was a period of continental accretion, deformation, and magmatism between about 2,150 and 1,960 Ma, during which the Imatata and the greenstone-granite terranes were deformed and metamorphosed.

Volcanic, subvolcanic, and plutonic rocks of the Cuchivero Group represent postcollisional, post-Trans-Amazonian magmatism in the Guyanan Shield about 1,930–1,790 Ma. Silicic rocks (rhyolite and granite to granodiorite) dominate; intermediate to mafic dikes and lava flows are less abundant. Mining has not occurred in rocks of the Cuchivero Group, although precious-metal, tin, and molybdenum prospects are present, and Olympic Dam-type iron-copper-uranium-gold-rare earth element deposits are permissive. The only known diamond-bearing kimberlite deposit in the Guayana Shield is in the Quebrada Grande area. This deposit and the carbonatite at Cerro Impacto are within the outcrop area of the Cuchivero Group. The kimberlite was emplaced about 1,732±82 Ma; however, the carbonatite is not dated. Possible periods of intrusion are about 1.7 Ga or during the Mesozoic after the opening of the Atlantic Ocean.

Undivided Proterozoic rocks in Estado Amazonas of western Venezuela include granitic rocks, gneiss, and migmatite. Metamorphism and magmatism were most intense in this area about 1,860–1,730 Ma.

Unmetamorphosed, posttectonic sedimentary rocks including quartzaronite, conglomerate, arkose, siltstone, and shale of the Roraima Group were deposited in fluvial,

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deltaic, shallow-marine, and lacustrine or epicontinental environments. Some rocks of the Roraima Group are at least 1,670 Ma in age. and others are possibly as old as about 1,900 Ma and as young as about 1,500 Ma or younger. Paleoplacer deposits of gold and diamonds in the lower part of the Roraima Group are the source for modern placers.

Continental-type tholeiitic dikes, sills, and irregular intrusive bodies of the Avanavero Suite cut all older rocks of the Guayana Shield. These intrusions are about 1,650 Ma in age and may be as old as about 1,850 Ma.

Middle Proterozoic (about 1.55 Ga) undeformed granite having rapikivi texture is characteristic of the Parguaza province. Quartz veins, pegmatite, and greisen in the Parguaza Granite have moderate potential for tin deposits. Similarities in age, composition, and tectonic environment indicate that Olympic Dam-type iron-copper-uranium-gold-rare earth element deposits may be present in the Parguaza Granite and its associated volcanic rocks.

Continental collision in the westernmost part of the Guayana Shield during the Nickerie orogeny reset many potassium-argon and rubidium-strontium mineral ages of Archean and Early Proterozoic rocks in the central and eastern parts of the shield to about 1,200 Ma. Tholeiitic diabase dikes intruded the Guayana Shield during the opening of the Atlantic Ocean from about 210 to 200 Ma. Variable lithologic resistance to weathering and erosion of a thick sequence of flat-lying to very gently dipping sedimentary strata in the Roraima Group have produced at least six planation surfaces in the Guayana Shield at distinct elevations between about 2,900 and 50 m above sea level.

Tropical weathering of the diverse rock types of the Guayana Shield has formed many deposits of bauxite and lateritic bauxite. The largest bauxite deposit is Los Pijiguaos, which developed on the Parguaza Granite. Deposits of bauxite and enriched banded iron formation formed on the Imataca-Nuria erosional surface. Placer diamond and gold deposits are mined in modern channels of the major rivers and in colluvial-alluvial deposits in low-order drainages.

Rocks, mineral deposits, and tectonic events in the Venezuelan Guayana Shield are generally correlative with those elsewhere in the Guayana Shield and (or) in the West African craton. Although Archean rocks are not known elsewhere in the Guayana Shield, Archean rocks similar to those in the Imataca Complex and coeval Archean tectonic events are present in the West African craton. Rocks in the Early Proterozoic greenstone belts in Venezuela are comparable in age and lithology to greenstone-belt rocks throughout the Guayana Shield and in the West African craton. The Trans-Amazonian orogeny in northern South America and the Eburnean orogeny in West Africa are the major Early Proterozoic tectonic event. Differences in the types of late Early Proterozoic and Middle Proterozoic rocks in the Guayana Shield and the West African craton are significant. For example, granite and rhyolite similar to the Cuchivero Group are present throughout the Guayana Shield, but are only locally present in the West African Shield. Also, rocks comparable to the Roraima Group and the Parguaza Granite, although widespread in the Guayana Shield, are rare or absent in West Africa.

RESUMEN

El Escudo de Guayana en Venezuela está compuesto por 5 provincias litotectónicas: (1) un terrano Arqueano con metamorfiamo de la facie de la anfibolita a la granulita, (2) un terrano granítico-cinturones de rocas verdes de edad Proterozoio Temprano, (3) un complete volcánico-plutónico sin metamorfismo, de edad Proterozoico Temprano, (4) rocas continentales de edad Proterozoico Temprano, y (5) granito rapakivi anorogénico de edad Proterozoico Medio. Las rocas del Proterozoico Temprano en el Estado Amazonas no han sido diferenciadas, por consiguiente, su relación con otras rocas del Escudo de Guayana en Venezuela no ha sido definida. Los diques, sills y cuerpos intrusivos pequeños e irregulares, de composición toleítica y origen continental de edad Proterozoico Temprano a Medio, y diques de edad Mesozoico que fueron emplazados durante la apertura del Oceano Atlántico, cortan las rocas de todas las provincias litotectónicas. Los principales yacimientos minerales son de oro, hierro, bauxita y diamantes.

El Complejo de Imataca de edad Arqueano es la unidad más antigua del escudo, consiste de gneises y granulita con cantidades menores de formación bandeada de hierro y dolomita. El estilo estructural consiste de grandes pliegues isoclinales los cuales han sido replegados formando pliegues relativamente más abiertos. El grado metamórfico varía desde la facie de la granulita, en la parte noreste del cinturón, a la facie de la anfibolita en la parte suroeste. Los depósitos de hierro en formaciones bandeadas de hierro enriquecidas contienen aproximadamente 2 billones de toneladas métricas de mineral de hierro. El Complejo de Imataca fue afectado durante el evento tectono-magmático pre-Trans-Amazónico, entre 2,800 y 2,700 Ma, cuando se formaron gneis de inyección y migmatitas.

Los cinturones de rocas verdes del Proterozoico Temprano se formaron aproximadamente entre 2,250 y 2,100 Ma. Estos consisten de una secuencia submarina de rocas volcánicas máficas toleíticas, una secuencia toleítica a calco-alcalina de basalto a riolita y una secuencia de grauwacas turbidíticas, rocas volcaniclásticas y rocas sedimentarias de origen químico las cuales caracterizan la base, parte media y partes superiores respectivamente. También se encuentran complejos estratificados maficos-ultramáficos. El grado metamórfico varía entre las facies del esquisto verde y anfibolita. Las zonas de cizalla que cortan los cinturones de rocas verdes contienen numerosos depósitos de oro en vetas de cuarzo con bajo contenido de sulfuros.

Los cinturones de rocas verdes fueron intrusionados por domos graníticos del Complejo de Supamo aproximadamente entre 2,230 y 2,050 Ma. Los domos graníticos dividieron los cinturones de rocas verdes en sinclinorios los cuales, en vista en mapa, presentan ramificaciones. La orogenesis Trans-Amazónica fue un período de colisión y crecimiento continental, deformación y magmatismo entre 2,150 y 1,960 Ma durante el cual las rocas del Complejo de Imataca y la provincia granítico–rocas verdes fueron deformados y metamorfizados.

Las rocas volcánicas, subvolcánicas y plutónicas del Grupo Cuchivero representan un período de magmatismo entre 1,930 y 1,790 Ma después de la colisión Trans-Amazónica. Predominan rocas silícicas (riolita y granito a granodiorita); diques y lavas intermedias a máficas son menos abundantes. Aunque existen prospectos de minerales preciosos, estaño y molibdeno, y prospectos del tipo Olympic Dam cobre-uranio-oro estan permisivo, hasta la fecha no hay operaciones mineras en rocas del Grupo Cuchivero. La única kimberlita diamantífera conocida en el Escudo de Guayana está localizada en el área de la Quebrada Grande. Este depósito, y la carbonatita del Cerro Impacto están localizados dentro de la región del Grupo Cuchivero. La kimberlita fue intrusionada aproximadamente hace 1,732±82 Ma. Aunque la edad de la carbonatita no ha sido definida, las edades posibles de intrusión son acerca de 1.7 Ma o durante el Mesozoico después de la apertura del Océano Atlántico.

Las rocas de edad Proterozoico en el Estado Amazonas no han sido divididas, pero incluyen rocas graníticas, gneis y migmatita. En esta área, el magmatismo y metamorfismo fueron más intensos aproximadamente entre 1,860 y 1,730 Ma.

Las rocas sedimentarias del Grupo Roraima fueron depositadas después del tectonismo Trans-Amazónico en ambientes fluviales, deltáicos, marino somero y lacustrino o epicontinentales. Estas rocas no están metamorfizadas e incluyen cuarzo-arenita, conglomerado, arcosa, limolita y lutita. Algunas de las rocas del Grupo Roraima tienen una edad de al menos 1,670 Ma, otras posiblemente tienen una edad de 1,900 Ma y otras pueden ser tan jovenes como 1,500 Ma. Los depósitos de oro y diamantes en paleoplaceres en la parte inferior del Grupo Roraima son la fuente para depósitos de placer modernos.

La Suite Avanavero, la cual comprende diques, sills y cuerpos intrusivos irregulares de composición toleítica continental, cortan todas las unidades más antiguas del Escudo de Guayana. Estas intrusiones probablemente se formaron aproximadaments hace 1,650 Ma, y quizás hace 1,850 Ma.

El granito rapakivi, no deformado, de edad Proterozoico Medio (aproximadamente 1,545 Ma), conforma la Provincia de Parguaza. El granito de Parguaza presenta vetas de cuarzo, y greisen conteniendo un potencial moderado para depósitos de estaño. La similitud en edad, composición y ambiente tectónico del granito de Parguaza con las rocas que contienen el depósito Olympic Dam de hierro, cobre, uranio, oro y elementos tierras raras, indican que este granito y rocas volcánicas asociadas tienen potencial para esos minerales.

La colisión continental en el extremo occidental del Escudo de Guayana, ocurrida durante la orogenesis Nickerie aproximadamente hace 1,200 Ma, re-estableció las edades isotópicas de los sistemas potasio-argón y rubidio-stroncio de las rocas Arquenas y Proterozoico presentes en la parte central y oriental del escudo. Durante la apertura del Océano Atlántico, entre 210 y 200 Ma, diques de diabasa toleítica intrusionaron el Escudo de Guayana. El levantamiento y erosión del los terrenos Precámbricos durante las Eras Mesozoica y Cenozoica produjeron al menos seis superficies de erosión en el Escudo de Guayana a distintas elevaciones entre 2,900 y 50 m.s.n.m.

La meteorización en clima tropical, de los distintos tipos de rocas del Escudo de Guayana, ha formado muchos depósitos de bauxita y bauxita laterítica. El depósito más grande es Los Pijiguaos, el cual se formó sobre el Granito de Parguaza. También se formaron depósitos de bauxita y hierro enriquecido a partir de formación bandeada de hierro en la superficie de erosión Imataca-Nuria. Diamantes y oro en depósitos de tipo placer son actualmente minados en canales activos de los ríos mayores y también en depósitos coluvio-aluviales en drenajes de orden menor.

Las rocas, depositos minerales y eventos tectónicos en el Escudo de Guayana en Venezuela son generalmente correlacionables con esos en otras partes del Escudo de Guayana y/o en el Craton de Africa Occidental. Aunque no se conocen rocas Arqueanas en otras partes del Escudo de Guayana, las rocas y eventos tectónicos en Africa Occidental de edad Arqueana son similares a las del Complejo de Imataca. Las rocas Proterozoicas de los cinturones de rocas verdes "greenstones" en Venezuela son comparables en edad y litología al resto de las rocas en los cinturones de rocas verdes a traves del Escudo de Guayana y en el Craton de Africa Occidental. La orogenia Trans-Amazónica ocurrida en la parte norte de Suramerica, y la orogenia Eburneana en Africa Occidental es el evento tectónico más importante durante el Proterozoico Temprano. En contraste a la similitud entre rocas Arqueanas y del Proterozoico Temprano entre Suramerica y Africa, existen diferencias significativas entre los tipos de rocas del Proterozoico Temprano tardío y Proterozoico Medio entre las mencionadas regiones. Por ejemplo, a traves del Escudo de Guayana hay granitos y riolitas similares a aquellos del Grupo Cuchivero, mientras que en el Escudo de Africa Occidental estas rocas están presentes solo localmente. Lo mismo sucede con las rocas del Grupo Roraima y el granito de Parguaza, las cuales afloran a traves del Escudo de Guayana pero no así en Africa Occidental donde raramente afloran o están ausentes.

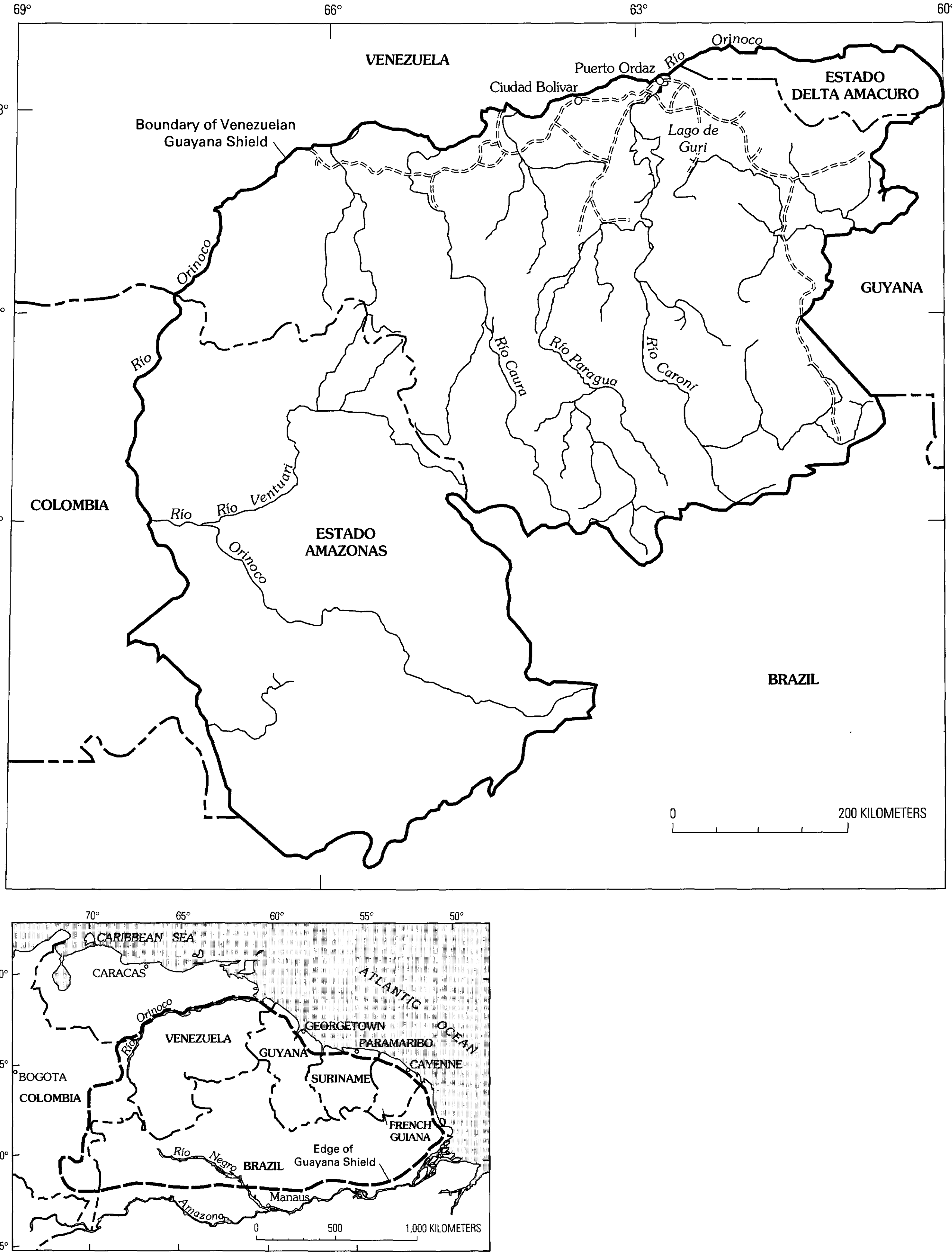

Figure 1. Major geographic features of the Venezuelan Guayana Shield. Smaller map shows outline of Guayana Shield of northeastern South America (modified from Gibbs and Barron, 1983).

INTRODUCTION

The U.S. Geological Survey assisted the Corporación Venezolana de Guayana, Técnica Minera, C.A. (CVG–TEC MIN, or TECMIN) between 1987 and 1992 in its assessment of and exploration for new mineral deposits in the Precambrian Guayana Shield of Venezuela (Wynn, Sidder, and others, this volume). The Guayana Shield, in the northern part of the Amazonian craton of South America, measures about 1,100 km north to south and 2,100 km east to west, covering an area of about 2,310,000 km^2 (fig. 1). Shield rocks crop out in Colombia, Venezuela, Guyana, Brazil, Suriname, and French Guiana. The Guaporé or western Central Brazil Shield, south of the Amazon River Basin, forms the southern part of the Amazonian craton (Gibbs and Barron, 1983; Teixeira and others, 1989; Goodwin, 1991). Although now separated by the Atlantic Ocean, the geologic histories of the Guayana Shield and the West African craton indicate that Archean and Early Proterozoic rocks of both areas were deposited in similar environments and were affected by coeval tectonic and metamorphic events.

The Guayana Shield in Venezuela consists of five lithotectonic provinces: (1) Archean amphibolite- to two-pyroxene granulite-facies gneiss terrane, (2) Early Proterozoic greenstone-granite terrane(s), (3) Early Proterozoic unmetamorphosed volcanic-plutonic complex, (4) Early to Middle Proterozoic continental sedimentary rocks, and (5) Middle Proterozoic anorogenic rapakivi-type granite (Gibbs and Barron, 1983; Teixeira and others, 1989). Early Proterozoic and possibly Archean rocks in Estado Amazonas of western Venezuela are undivided, and their relation to other rocks of the Venezuelan Guayana Shield is uncertain. Early to Middle Proterozoic continental-type tholeiitic dikes, sills, and small irregular intrusive bodies and Mesozoic dikes emplaced during the opening of the Atlantic Ocean are present in all of the lithotectonic provinces. Table 1 is a simplified stratigraphic chart of the rock units and tectonic events of the Guayana Shield of Venezuela, and figure 2 is a simplified geologic province map. Geologic and geographic maps of the Guayana Shield of Venezuela are presented in U.S. Geological Survey and Corporación Venezolana de Guayana, Técnica Minera, C.A. (1993), and mineral deposits mentioned herein are shown on plate 1 of Sidder (this volume).

In this paper, we present an overview of the geology of the Venezuelan Guayana Shield and discuss its relation to the geology of the entire Guayana Shield. The discussion of the evolution of the shield includes recently published uranium-lead and samarium-neodymium isotopic data. Geochronological data have been standardized by recalculation with one set of constants as recommended by the Subcommission on Geochronology of the International Union of Geological Sciences[3] (Steiger and Jäger, 1977); errors in all dates are given at the 1-sigma level. Such standardization helps to define more narrowly the ranges of specific magmatic or tectonic events. For example, a single tectonic episode such as the Trans-Amazonian orogeny, which previously had been reported to span 400 m.y., is herein constrained to a much shorter interval of 190 m.y. Similarly, the formation of the Supamo Complex (granite and gneiss associated with the greenstone-belt rocks), which reportedly occurred between 2,700 and 2,100 Ma, is herein restricted to 2,230–2,050 Ma. These recalculated dates of the major geologic events in the Venezuelan Guayana Shield allow its history to be interpreted more realistically (table 1, fig. 2).

IMATACA COMPLEX

The Archean Imataca Complex is a northeast-trending belt of amphibolite- to granulite-facies metasedimentary and metaigneous rocks. This belt is at least 510 km long and 65–130 km wide, and it forms the northernmost margin of the Venezuelan Guayana Shield (fig. 2). Rocks of the Pliocene and Pleistocene Mesa Formation (not shown in fig. 2) and alluvium from the floodplain of the Río Orinoco cover the Imataca Complex along its northern margin, and the Guri shear zone separates the Imataca Complex from the Early Proterozoic greenstone-granite terrane to the south. The Imataca Complex abuts against plutonic and volcanic rocks of the Early Proterozoic Cuchivero Group along the Río Caura in the west. The nature of the contact is unknown because it is obscured by thick overburden and alluvium along the Río Caura (Kalliokoski, 1965; Ascanio, 1975; Mendoza, 1977a).

The Imataca Complex includes more than 80 percent quartzofeldspathic orthogneiss, paragneiss, and felsic granulite, 10–15 percent intermediate to mafic orthogneiss, granulite, and charnockite, 1 percent metamorphosed banded iron formation, and minor manganiferous metasedimentary rocks, dolomitic marble, and anorthosite. The protolith of the Imataca Complex consisted of clastic and chemical sedimentary rocks, silicic calc-alkaline subaerial volcanic rocks, and lesser plutonic rocks (Kalliokoski, 1965; Dougan, 1977; Gibbs and Wirth, 1986).

The grade of metamorphism in the Imataca Complex varies from two-pyroxene granulite facies in that part of the belt generally northeast of the Lago de Guri area to

[3]All rubidium-strontium isochron dates reported here have been recalculated with the decay constants and isotopic abundances recommended by Steiger and Jäger (1977): ^{87}Rb decay constant=1.42×10^{-1} yr^{-1}; atomic ratio ^{85}Rb/^{87}Rb=2.59265; atomic ratio ^{86}Sr/^{88}Sr=0.1194; atomic ratio ^{84}Sr/^{86}Sr=0.056584. A best-fit line has been calculated by the method of York (1969). Dates reported are those from model 3 of York, which assumes that the scatter of data is due to a combination of the assigned analytical error and a normally distributed variation in the initial ^{87}Sr/^{86}Sr. All potassium-argon dates have been recalculated (where sufficient data are available) using the decay constants and isotopic abundances recommended by Steiger and Jäger (1977): $\lambda^{40}K_e+\lambda^{40}K_e=0.581\times10^{-10}$/yr; $^{40}K_\beta=4.962\times10^{-10}yr^{-1}$; ^{40}K=0.01167 atomic percent (1.167×10^{-4} mol/mol); or by conversion with the critical table for conversion of K-Ar ages from old western constants to new IUGS constants (Dalrymple, 1979).

Table 1. Rock units and tectonic events of the Guayana Shield of Venezuela.
[Rock unit designations as used in figure 2]

Alluvium (Qal)—Quaternary alluvial sediments
MESOZOIC-CENOZOIC UPLIFT (uplift, tilt, and formation of erosion surfaces)
Diabase dikes (Mzd)—Thin, elongated tholeiitic dikes (about 210–200 Ma)
NICKERIE OROGENY (about 1,200 Ma)
Parguaza Granite (Yp)—Massive, coarsely crystalline, porphyritic granite and biotite granite, commonly with rapakivi (wiborgite-type) texture (about 1,550 Ma)
Avanavero Suite (Xa)—Continental tholeiitic dikes, sills, inclined sheets, and small irregular intrusive bodies (about 1,650 Ma; possibly as old as 1,850 Ma)
Roraima Group (YXr)—Continental (fluvatile-deltaic and lacustrine) quartz sandstone and quartz-pebble conglomerate with lesser feldspathic arenite, arkose, siltstone, shale, jasper, chert, and interlayered felsic volcanic rocks (possibly as young as about 1,500 Ma; possibly as old as about 1,900 Ma; dates from Uaimapué Formation are older than 1,650 Ma). Includes the Matauí Formation, the Uaimapué Formation, the Kukenán Formation, and the Uairén Formation
Undivided Proterozoic rocks (Pu)—Synkinematic plutonic rocks (granite to tonalite and quartz diorite) and medium- to high-grade gneiss with both igneous and sedimentary protoliths in the Estado Amazonas only (about 1,860–1,730 Ma)
UNNAMED OROGENY (Estado Amazonas; about 1,860–1,730 Ma)
Cuchivero Group (Xc)—Thick sequence of unmetamorphosed felsic to intermediate subaerial volcanic rocks and their associated granitic rocks (about 1,930–1,790 Ma). Includes the granite of Guaniamito, the granites of San Pedro and Santa Rosalía (including the granite of Las Trincheras), and the Caicara Formation
TRANS-AMAZONIAN OROGENY (about 2,150–1,960 Ma)
Supamo Complex (Xs)—Gneiss, schist, migmatite, and granitic rocks such as trondhjemite (sodic granite), tonalite, granodiorite, and quartz monzonite associated with the greenstone-belt terrane (2,230–2,050 Ma)
Greenstone-belt rocks (Xg)—Sequences as thick as 11,000 m of metamorphosed tholeiitic basalt and gabbro with interflow chemical sedimentary rocks in the lower part of the sequence; interstratified, porphyritic, tholeiitic and calc-alkalic basaltic to rhyolitic lava flows and tuffs in the middle part; and tuffaceous, volcaniclastic, turbiditic, pelitic, and chemical sedimentary rocks in the upper part. About 2,250–2,100 Ma. Includes the Los Caribes and Caballape Formations of the Botanamo Group; the Yuruari Formation and the Cicapra and El Callao Formations of the Pastora Supergroup; and the Real Corona–El Torno assemblage
PRE-TRANS-AMAZONIAN TECTONOMAGMATIC EVENT (about 2,800–2,700 Ma)
Imataca Complex (Ai)—Amphibolite- to granulite-facies quartzofeldspathic orthogneiss, paragneiss, and felsic granulite, intermediate to mafic orthogneiss, granulite, and charnockite, metamorphosed banded iron formation, and minor manganiferous metasedimentary rocks, dolomitic marble, and anorthosite (>2,800 Ma; protolith possibly 3,700–3,400 Ma)

amphibolite facies southwest of the area. The gneiss is commonly migmatitic and consists of quartz-potassium feldspar-plagioclase$\pm$biotite$\pm$hornblende$\pm$orthopyroxene$\pm$clinopyroxene$\pm$garnet$\pm$sillimanite$\pm$cordierite$\pm$muscovite. Estimates of peak metamorphic conditions in granulite-facies rocks indicate that temperature was between about 750°C and 800°C and pressure was between about 8.0 and 8.5 kb (Short and Steenken, 1962; Swapp and Onstott, 1989); in the amphibolite-facies rocks, temperature was between about 625°C and 700°C, and pressure was between 4 and 7 kb (Dougan, 1974, 1977).

Rocks of the Imataca Complex are strongly deformed. The entire stratigraphic sequence of the complex was folded into large isoclinal folds that were refolded by relatively open folds (Ruckmick, 1963; Onstott and others, 1989). In the northern part of the Imataca Complex, the isoclinal fold axes strike generally northwest, whereas in the southern part they strike east-west. Fold axes are deflected to the northeast close to the Guri shear zone (fig. 2), which can be traced in the field for a distance of more than 400 km (Onstott and others, 1989). This shear zone, from several hundred meters to 1 km in width, is marked by alternating bands of mylonite, pseudotachylite, and strongly sheared gneiss and amphibolite. Crushed rock is visible in thin sections of samples collected as far away as 2 km on either side of this shear zone (Short and Steenken, 1962). The deflection of structural trends and mineral lineations adjacent to the Guri fault indicates that major movement on the fault was left-slip, possibly with later vertical displacement. The rocks of the Imataca Complex are also cut by north-verging low-angle thrust faults that are associated with the Guri shear zone and in part caused the folding (Ascanio, 1975). High-grade mylonite zones, such as the El Pao and the Río Claro fault zones, also cut the Imataca Complex (Short and Steenken, 1962; Onstott and others, 1989; Swapp and Onstott, 1989). These faults may have been reactivated one or more times in the Proterozoic and Phanerozoic during the Nickerie and Caribbean orogenies (Gibbs and Barron, 1993; Olmore and others, 1993). For example, faults that cut the Imataca Complex and are subparallel to the Guri fault zone show evidence of postuplift, Cenozoic, right-lateral and normal displacement (Olmore and others, 1993).

AGE

Most of the radiometric dates of rocks in the Imataca Complex record regional metamorphic and magmatic events. Metasedimentary protoliths for some gneissic rocks of the Imataca Complex have been dated at 3,700–3,400 Ma by whole-rock rubidium-strontium isochron and lead-lead methods (Montgomery, 1979); it is possible that these dates reflect an inherited detrital Archean component rather than a primary age of deposition (R.M. Tosdal, U.S. Geological Survey, oral commun., 1990). Rocks of the Imataca Complex were deformed, intruded, and regionally metamorphosed about 2,800–2,700 Ma. During the Trans-Amazonian orogeny, about 2,150–1,960 Ma, they underwent upper amphibolite- to granulite-facies metamorphism and granitic intrusion (Hurley and others, 1976; Onstott and

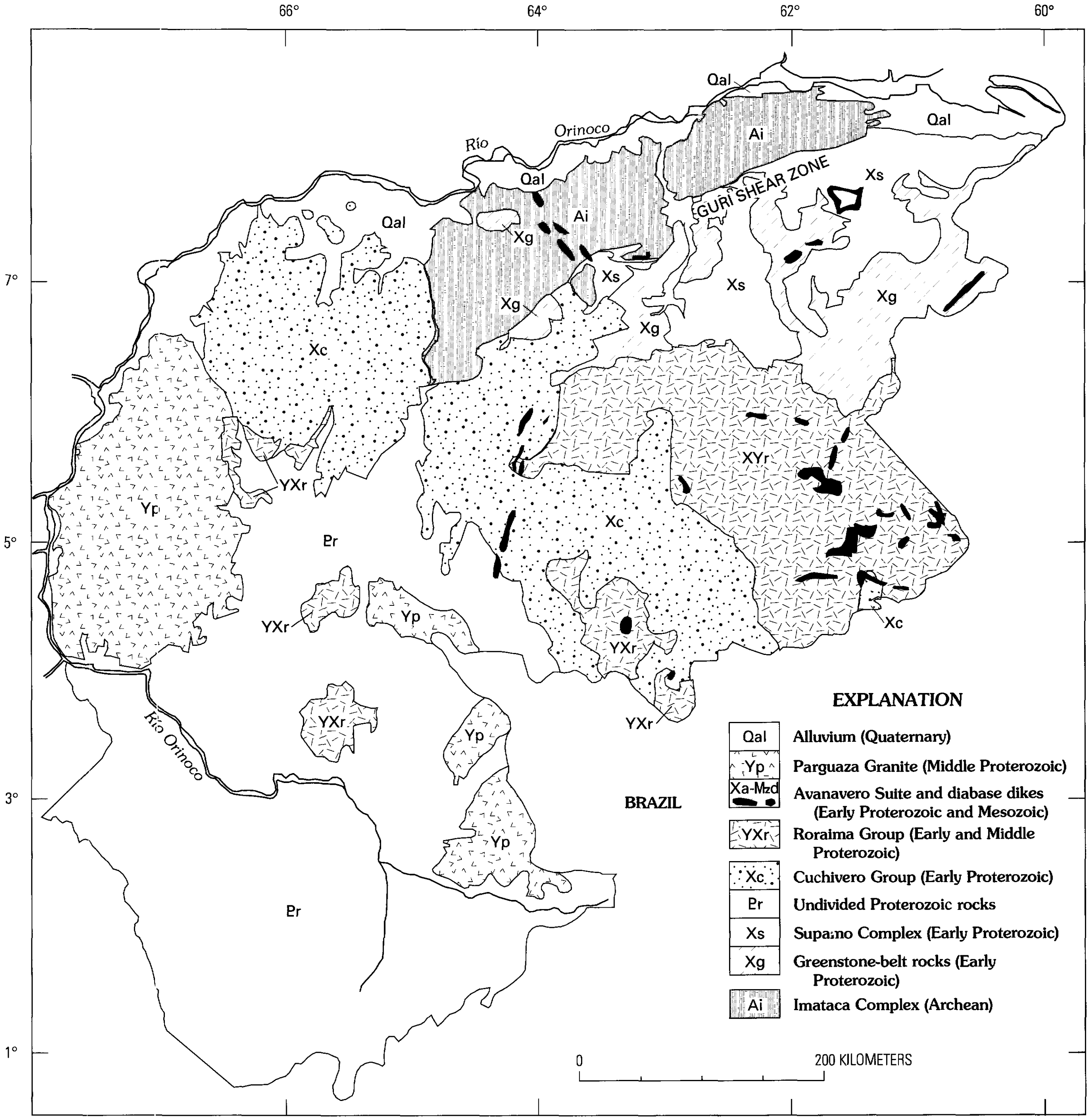

Figure 2. Simplified geologic provinces of the Venezuelan Guayana Shield. Rock units are described in table 1. Modified from Rodriguez and others (1976).

others, 1989). A sensitive high mass-resolution ion microprobe (SHRIMP) study of the uranium-lead ages of zircons collected from sand in the Río Orinoco west of Ciudad Bolívar identified a small, discrete population of zircons that have an age of about 2,800 Ma in addition to a larger population having an age of about 2,100–2,000 Ma (Goldstein and Arndt, 1988). The scarcity of zircon grains of Archean age demonstrates that Archean rocks form only a minor portion of the Venezuelan Guayana Shield (Goldstein and Arndt, 1988).

Rocks correlative in age to the Imataca Complex are not known elsewhere in the Guayana Shield; however, amphibolite- to granulite-facies rocks, including banded iron formation, in the Kenema-Man domain of the Leo Shield and in the western Reguibat Shield of the West African craton may be correlative with rocks of the Imataca Complex (Cahen and others, 1984; Cohen and Gibbs, 1989; Rocci and others, 1991). In the Archean terrane of West Africa, quartzofeldspathic gneiss, migmatite, and granite are ubiquitous, and layered and massive gabbro and ultramafic rocks, amphibolite, banded iron formation, anorthosite, and calc-silicate or marble beds are common (Williams, 1988; Rocci and others, 1991). The peak temperature of metamorphism for these West African shield rocks was about 750°C–850°C, and pressure estimates of 7–10 kb indicate a depth of equilibration of about 21–30 km (Williams, 1988; Rocci and others, 1991). Paleomagnetic reconstructions and geochronological data indicate that the Guri shear zone in the Venezuelan Guayana Shield is aligned with the Sassandra–Trou Mountain fault zone in the Leo Shield of the West African craton, which may also be equivalent to the Zednes fault in the Reguibat Shield. The latter two faults also separate Archean high-grade metamorphic rocks from Early Proterozoic, lower grade greenstone-granite terranes (Onstott and Hargraves, 1981; Caen-Vachette, 1988; Cohen and Gibbs, 1989; Rocci and others, 1991; Boher and others, 1992).

Several high-grade metamorphic terranes in the Guayana Shield and the West African craton have been considered to be Archean in age because of their intense deformation and high-grade metamorphism. For example, granulite and charnockite in the Apiaú Complex, Brazil, the Kanuku Complex, Guyana, and the Falawatra Group in the Bakhuis Mountains, Suriname, are part of the central Guyana granulite belt (Gibbs and Wirth, 1986; Gibbs and Barron, 1993). Although these rocks and the L'Ile de Cayenne Complex in French Guiana have structural, stratigraphic, and petrographic similarities to the Imataca Complex, they have apparent protolith ages of about 2,300–2,200 Ma, and peak metamorphism in these rocks is related to the Trans-Amazonian orogeny at about 2,000 Ma (Priem and others, 1978; Ben Othman and others, 1984; Teixeira and others, 1984; Gibbs and Wirth, 1986; Rowley and Pindell, 1989; Teixeira and others, 1989; Gibbs and Barron, 1993). Uranium-lead and rubidium-strontium isotopic evidence for an age older than 2,400 Ma does not exist for these rocks (Priem and others, 1978; Gibbs and Barron, 1993). Two quasicratonic nuclei, the Pakaraima nucleus in Estado Amazonas of Venezuela and the Xingu nucleus of northern Brazil, that consist of granitoid gneiss, migmatite, amphibolite, quartzite, and schist are assumed to be Archean in age because of their upper amphibolite- to granulite-facies metamorphic grade (Cordani and Brito Neves, 1982; Goodwin, 1991); however, only a few radiometric dates have been reported for these areas (Cordani and Brito Neves, 1982; Goodwin, 1991), and the age of these granitic and gneissic rocks is still unknown. Similar high-grade metaigneous and metasedimentary gneisses in the West African craton have precise isotopic ages of formation of about 2.19–2.14 Ga, and metamorphism occurred about 2.15–2.14 Ga (Boher and others, 1992). Tonalitic and trondhjemitic rocks associated with granulitic rocks in the Cupixi area, Amapa Federal Territory, northeastern Brazil, may be Late Archean in age (De Vletter and Kroonenberg, 1984; Teixeira and others, 1989).

MINERAL DEPOSITS

Iron is the predominant metal produced from the Imataca Complex. Several deposits of enriched banded iron formation in the Imataca Complex, such as Cerro Bolívar and San Isidro (Sidder, this volume, pl. 1), rank among the world's largest (Gruss, 1973; Sidder, this volume). Reserves of iron ore are greater than 1,855 million metric tons at a grade of about 63 percent iron and about 11,700 million metric tons at a grade of about 44 percent (Rodriguez, 1986, 1987; Doan, 1994). Banded iron formation protore consisted of an oxide facies assemblage in which magnetite and hematite were the dominant iron minerals. Enriched banded iron formation ore, composed predominantly of goethite and limonite, generally is in the limbs and centers of synclines. The iron-rich beds are intimately interbedded with layers of silica, present as quartz, and iron-bearing metamorphic minerals such as greenalite, grunerite, cummingtonite, crossite-magnesioriebeckite, acmite, and chlorite (Ruckmick, 1963; Gruss, 1973; Ascanio, 1985; Moreno and Bertani, 1985a). These deposits are most similar to Superior-type banded iron formation, although some Algoma-type banded iron formation may also be present (Sidder, this volume). The precious-metal content of these deposits is apparently low (Engineering and Mining Journal, 1987).

Small deposits and prospects of manganese and bauxite are present in the Upata–El Palmar–Guacuripia area (Sidder, this volume, pl. 1). Beds of secondarily enriched manganese ore are interstratified with gneiss, migmatite, amphibolite, and granulite of the Imataca Complex (Drovenik and others, 1967). These rocks are part of a stratigraphic sequence of gondite, quartz-biotite schist, amphibole schist, and dolomitic marble that is less than 500 m thick. The individual manganiferous beds are generally less than 10 m thick and have strike lengths of as much as 20 km or more (Drovenik and others, 1967). Drovenik and others concluded that the sedimentary-nonvolcanogenic manganese deposit model best represents the protore deposits of manganese in the Imataca Complex; however, the association with felsic to intermediate volcanic rocks in some areas (Dougan, 1977) rather than with a sedimentary protolith suggests that the sedimentary-volcanogenic manganese model (Sidder, 1991) may also characterize some of the manganese deposits. Bauxite in the Upata district is locally associated with weathered gabbro, amphibolite, and possibly granitic gneiss of the Imataca Complex (Candiales, 1961). A large-tonnage, low-grade high-silica bauxite deposit was recently discovered north of El Palmar on the Palsapa Plateau. Volcanogenic massive sulfide deposits have not been discovered in the Imataca Complex or elsewhere in the Guayana Shield (Gibbs and Barron, 1983, 1993).

PRE-TRANS-AMAZONIAN TECTONOMAGMATIC EVENT

Intrusion of homogeneous granitic rocks and injection gneiss and the development of migmatite characterize the pre-Trans-Amazonian tectonomagmatic event about 2,800–2,700 Ma in the Guayana Shield. The Cerro La Ceiba migmatite in the Imataca Complex is representative of rocks formed at this time (Hurley and others, 1976).

This event is correlative with the Liberian tectonothermal event in the Leo and Reguibat Shields of the West African craton, during which Archean rocks were metamorphosed and intruded by plutonic rocks between about 2,780 and 2,750 Ma (Hedge and others, 1975; Rollinson and Cliff, 1982; Tysdal and Thorman, 1983; Cahen and others, 1984). This event is also known as the Aroense in Venezuela, the Guriense in Guyana, and the Jequié in Brazil (Singh, 1974; Schobbenhaus and others, 1984).

GREENSTONE BELTS

Early Proterozoic greenstone-granite terranes in the central and eastern parts of the Venezuelan Guayana Shield (fig. 2) comprise an area of about 360 km by 250 km. Felsic to intermediate volcanic rocks of the Cuchivero Group and clastic sedimentary rocks of the Roraima Group overlie rocks of the greenstone-granite terranes to the west and south. Rocks of the greenstone-granite terranes are continuous with those identified in Guyana to the east. The contact between the greenstone-granite terranes and the Imataca Complex to the north is along the Guri shear zone (Gibbs and Olszewski, 1982; Onstott and others, 1989).

Rocks of the greenstone belts in Venezuela have a total thickness of 11,000 m or more. Named stratigraphic units of the Venezuelan greenstone belts include the Pastora Supergroup and the Botanamo Group (table 1). The Pastora Supergroup, which consists of the Carichapo Group (El Callao and Cicapra Formations) and the Yuruari Formation, is discordantly overlain by the Botanamo Group, which consists of the Caballape and Los Caribes Formations. Granitic plutons and domelike batholiths, gneiss, and migmatite of the Supamo Complex divide these metasedimentary and metaigneous rocks of the greenstone belts into branching synclinoria (Menendez, 1968, 1972; Benaim, 1972, 1974; Cox, Gray, and others, 1993). Metasedimentary and metavolcanic rocks of the Real Corona–El Torno assemblage to the west of the Río Aro (fig. 2) are tentatively correlated with those of the Pastora Supergroup (Kalliokoski, 1965). Rocks of the Pastora-Botanamo greenstone belts are correlative with those of the Barama-Mazaruni Supergroup in Guyana, the Marowijne Group in Suriname, the Paramaca Series (Orapu and Bonidoro Groups) in French Guiana, and the Vila Nova Group in Brazil (Bosma and others, 1983; Gibbs and Barron, 1983, 1993; Schobbenhaus and others, 1984; Teixeira and others, 1984, 1989; Gruau and others, 1985). All of these metaigneous and metasedimentary rocks form part of the Maroni-Itacaiunas province, a so-called mobile belt or greenstone belt that makes up a large part of the supracrustal rocks of the Amazonian craton (Cordani and Brito Neves, 1982; Teixeira and others, 1989; Goodwin, 1991).

Early Proterozoic greenstone belts of the Guayana Shield are comparable in age and lithology to Birimian greenstone belts of the West African craton (Black, 1980; Kesse, 1985; Cohen and Gibbs, 1989; Boher and others, 1992; Sylvester and Attoh, 1992). Rocks of these greenstone belts have been correlated on the basis of grossly similar lithostratigraphy, chemistry, age, structure, and intensity of metamorphism. The rocks of the Guayana Shield and the West African craton probably represent several penecontemporaneous greenstone belts, not one continuous belt.

Rocks of the greenstone belts of the Guayana Shield were deposited predominantly in a submarine environment. Basalt containing pillow structures and chemical and mineralogical alteration characteristic of submarine spilitization dominates the lower parts of the greenstone-belt sequence. The middle part of the sequence has a higher proportion of porphyritic andesite, dacite, and rhyolite submarine and possibly subaerial lava flows and siliceous and tuffaceous interflow sediments. Turbiditic graywacke, pelite, tuff, chemical sedimentary rocks, and volcaniclastic rocks are dominant in the uppermost part of the greenstone-belt sequence. The transition from volcanic to sedimentary rocks is locally conformable (Menendez, 1972; Bosma and others, 1983; Gibbs and others, 1984; Gibbs and Wirth, 1986; Gibbs, 1987; Day and others, 1989). In the greenstone belt of western Guyana, which is continuous into Venezuela, basalt and gabbro (some representing slow-cooled interiors of thick flows and sills) make up about 75 percent of the igneous rocks, basaltic andesite and andesite flow rocks about 17 percent, and rhyolite flow and pyroclastic rocks about 8 percent (Renner and Gibbs, 1987). Both tholeiitic and calc-alkaline chemical trends are present in the volcanic rocks of the greenstone belts (Renner and Gibbs, 1987; Day and others, 1989).

Ultramafic rocks make up about 1–2 percent of the igneous rocks in the greenstone belts of the Guayana Shield (Gibbs, 1987). Komatiite has been tentatively identified in two isolated areas of the Guayana Shield in Venezuela on the basis of high magnesian content (22 weight percent MgO or more) (Tosiani and Sifontes, 1989); however, reported relic spinifex texture (Tosiani and Sifontes, 1989) is actually rosettes of amphibole after pyroxene in contact metamorphic aureoles of gabbroic intrusive rocks (Gray and others, this volume). Spinifex textures have not been reported elsewhere in the Guayana Shield (Gibbs, 1987), although peridotitic komatiite has been identified in central French Guiana on the basis of whole-rock chemistry (Gruau and others, 1985).

Mafic-ultramafic intrusive rocks are present throughout the stratigraphic sequence of the greenstone belts of the

Guayana Shield. They commonly form layered complexes that include cumulate rocks such as pyroxenite[4] and peridotite associated with gabbro and lesser anorthosite and diorite. These intrusive complexes are present both as strongly metamorphosed and deformed bodies and as relatively unmetamorphosed and undeformed bodies. In a few cases, the lower grade of metamorphism and less intense deformation are only in the interior parts of massive gabbroic bodies; the outer parts of these bodies are strongly deformed and metamorphosed. Some gabbroic rocks are relatively fresh and undeformed and therefore are apparently younger than the Trans-Amazonian orogeny (table 1); these gabbroic rocks may be associated with the Early to Middle Proterozoic Avanavero Suite (Benaim, 1972; Menendez, 1972, 1974; De Roever and Bosma, 1975; Gibbs, 1986; Gibbs and Barron, 1993). Wynn, Page, and others (this volume) describe a mafic-ultramafic layered complex at Pistón de Uroy, and Gray and others (this volume) describe similar rocks in the Sierra Verdún–Cerro Piedra del Supamo area.

Mafic and ultramafic rocks in the Venezuelan Guayana Shield form belts of small bodies of serpentinite and amphibole-talc-serpentine-carbonate rocks. Some of these gabbroic complexes apparently were preferentially intruded into the upper part of the El Callao Formation, where they are parallel to subparallel with basaltic lava flows of this formation. The maximum thickness of these gabbroic bodies is about 500 m (Menendez, 1972). Possibly some gabbro represents the slowly cooled interior of thick lava flows.

PASTORA SUPERGROUP

The Pastora Supergroup consists of the Carichapo Group (El Callao and Cicapra Formations) and the Yuruari Formation (table 1). The El Callao Formation is the oldest unit of the Pastora Supergroup. Its basal contact is everywhere intruded by granitic rocks of the Supamo Complex, and its upper contact is transitional with the Cicapra Formation or concordant with the Yuruari Formation. The El Callao Formation was originally described in the El Callao area by Korol (1965) and Menendez (1968). Rocks called El Callao Formation in other areas do not necessarily conform to the description of the unit from the type section. This is also true for other units of the greenstone belts; that is, descriptions outside of the area of the type section may not agree with those of the type section. Thus, it is possible, or even likely, that more than one greenstone belt is present in the Venezuelan Guayana Shield, even though present stratigraphic nomenclature implies that only one belt exists. Recent compilations of the geology of the Venezuelan Guayana Shield (Cox, Gray, and others, 1993; Cox, Wynn, and others, 1993) retain formation names only in the El Callao area. In other areas, rocks are divided on a lithologic basis into units such as mafic to intermediate rocks containing abundant chlorite and actinolite, felsic metavolcanic rocks containing abundant quartz and white mica, gabbro, peridotite, and mica schist and phyllite (Cox, Gray, and others, 1993; Cox, Wynn, and others, 1993).

The El Callao Formation, as thick as 3,000 m, consists almost exclusively of metamorphosed low-potassium basaltic to andesitic lava flows that commonly contain pillow structures and have amygdular and brecciated flow tops (Menendez, 1968, 1972; Benaim, 1972). Minor ferruginous quartzite and ferruginous and manganiferous chert (metamorphosed banded iron formation?) and talc schist lenses are present in several areas. Rocks of the El Callao Formation have been metamorphosed to the greenschist facies and locally to the almandine-amphibolite subfacies of the amphibolite facies. Greenschist-facies rocks typically are biotite-chlorite-albite-epidote±actinolite schist. Close to granitic intrusions of the Supamo Complex, the rocks are amphibolite containing blue-green hornblende and plagioclase (albite to andesine). A metamorphic and color zonation has been identified in pillow lavas as far as 6 km from the intrusive rocks (Menendez, 1968). The light-green greenschist-facies lavas are darker, grayish-green to greenish-black, amphibolite-facies rocks toward the intrusive bodies. Amphibolite is also abundant within 30 km of the Guri fault (Cox, Wynn, and others, 1993).

Hills having irregular crests typify the topographic expression of the El Callao Formation. They are 300–800 m in elevation, about 100–500 m above the surrounding terrain. In contrast, gabbro in mafic complexes forms slightly higher and smoother crested hills. Red soil on the El Callao Formation supports a dense forest (Menendez, 1968).

The Cicapra Formation overlies the El Callao Formation and includes a sequence as thick as 2,000 m of rhythmically bedded submarine andesitic tuff, turbiditic graywacke, and siltstone in packets about 10 m thick. Lithic tuff, tuff breccia, volcanic agglomerate, and, in the uppermost part of the formation, manganiferous hematitic chert are minor components of the Cicapra Formation (Menendez, 1972). These rocks are greenschist-facies, porphyroblastic, quartz-poor actinolite-biotite-epidote-albite schist. Amphibolite developed locally in this formation near granitic rocks does not contain biotite or porphyroblasts of amphibole. Schistosity, oblique to parallel to stratification, is generally poorly developed; in areas near granitic intrusive rocks, it is better developed. Rocks of the Cicapra Formation form a completely flat topography covered by a clayey soil the color of red wine. This unit wedges out and disappears southeast of El Callao, at which point the Yuruari Formation rests on the El Callao Formation (Menendez, 1968).

[4]For simplicity, the metaigneous and metasedimentary rocks in the greenstone belts are referred to by their precursor rock type name; that is, pyroxenite rather than metapyroxenite, basalt instead of metabasalt, graywacke instead of metagraywacke.

The basal contact between the Yuruari and the Cicapra Formations is gradational and that between the Yuruari and El Callao Formations is both depositional and faulted. The Yuruari Formation consists of mica schist, phyllite, and felsic metatuff possibly derived from epiclastic and turbiditic rocks; rhythmically bedded packets of feldspathic sandstone, siltstone, and black shale are as thick as 50 m. Locally, tuffaceous breccia, manganiferous phyllite, and intercalated dacitic to basaltic tuff, breccia, and lava flows, and chert are also present (Menendez, 1968, 1972; Benaim, 1972). Some phyllite and schist in the Yuruari Formation classified previously as metasedimentary rocks commonly contain relict phenocrysts of subhedral feldspar and rounded, embayed quartz, that indicates some phyllite and schist originated as felsic volcanic rocks (Cox, Gray, and others, 1993). The overall thickness of this formation is about 1,000 m (Menendez, 1968, 1972; Benaim, 1972; Day and others, this volume). Rocks of the Yuruari Formation are typically greenschist-facies chlorite-sericite±calcite schist. Only locally are rocks of the Yuruari Formation intruded by granite of the Supamo Complex. In aureoles of granitic intrusions, hornblende-hornfels and pyroxene-hornfels facies containing biotite, sillimanite or andalusite, chloritoid, tourmaline, and garnet developed. The upper contact does not crop out, but the contact with the overlying Caballape Formation is apparently an angular unconformity and (or) a tectonic disconformity (Menendez, 1968). Low hills and plains exhibiting a rectangular drainage pattern and a varicolored (light to dark yellow, reddish yellow, and several shades of red) clayey residual soil have developed on the Yuruari Formation. Savanna-type vegetation is characteristic (Benaim, 1972; Menendez, 1972).

Rocks of the greenstone belts of the Pastora Supergroup are strongly deformed and record at least two episodes of deformation. Recumbent isoclinal folds with folded axial planes are characteristic, and the axial planes are commonly parallel to subparallel with the borders of the granitic intrusions of the Supamo Complex (Menendez, 1972). Thus, the rocks of the Pastora Supergroup typically form synforms that wrap around granitic domes of the Supamo Complex. Foliation is commonly developed in rocks of the greenstone belt and is parallel to subparallel with the primary stratification. Foliation is best developed close to the contacts with the Supamo Complex. Cleavage parallel with the axial plane of the folds is also well developed in these rocks. Major shear zones as wide as 1 km and as long as 35 km cut the rocks of the Pastora Supergroup (Menendez, 1972, 1974).

BOTANAMO GROUP

Rocks in the Caballape and Los Caribes Formations of the Botanamo Group discordantly overlie the Pastora Supergroup. The Caballape Formation includes mafic to felsic lava and pyroclastic flows and breccia interbedded with epiclastic and turbiditic sedimentary rocks. Menendez (1968) estimated that graywacke, conglomerate, and siltstone make up 80 percent of the unit in the El Callao–Guasipati area, the remainder being andesitic to rhyodacitic pyroclastic tuff and breccia. Benaim (1972) noted, however, that only the basal part of the formation crops out in this area. Day and others (1989) determined that in the Anacoco area the Caballape Formation consists of about 80 percent basaltic to dacitic volcanic flows (some with pillow lavas) and associated pyroclastic rocks and about 20 percent volcanic breccia and graywacke and thin (1–5 cm thick) horizons of shale. Use of the name Caballape Formation should be restricted to those rocks in the type section area near El Callao. The basal contact of the Caballape Formation is discordant to unconformable with the Pastora Supergroup, and the upper contact is reportedly concordant with the Los Caribes Formation (Benaim, 1972). The minimum thickness of this formation is about 5,000 m (Menendez, 1968). Granite of the Supamo Complex apparently does not intrude rocks of the Caballape Formation, and thus amphibolite is not present in this formation, in contrast to the Pastora Supergroup (Cox, Wynn, and others, 1993). The Caballape Formation consists only of greenschist-facies schist containing chlorite, epidote, sericite, quartz, calcite, biotite, and opaque oxide minerals; the schist is only moderately folded into broad synforms. The terrain underlain by this formation is flat; low hills are elongate parallel with the trend of the beds, and the drainage pattern is rectangular or dendritic. The rocks weather to form a bleached soil (Benaim, 1972; Menendez, 1972).

The Los Caribes Formation consists of intercalated red phyllite and sandstone, polymict conglomerate, and siltstone and minor felsic tuff. Some authors excluded this unit from the greenstone-belt sequence and referred to it as pre-Roraima metasedimentary rocks or pre-Roraima foliated sandstone (Ghosh, 1985). It is likely that rocks of both pre-Cuchivero and post-Cuchivero age have been included within the Los Caribes Formation because some rocks mapped as Los Caribes Formation contain fragments of unmetamorphosed felsic tuff typical of the Caicara Formation of the Cuchivero Group, not greenschist-facies metavolcanic rocks (Cox, Wynn, and others, 1993). Thus, use of the name Los Caribes Formation should be restricted to that area near the type section in the Río Cuyuni where the contact between the Caballape and Los Caribes Formations has been mapped as concordant and interdigitated (Benaim, 1972).

Sedimentary rocks of the Los Caribes Formation are metamorphosed to the greenschist facies, whereas sedimentary rocks of the Roraima Group are unmetamorphosed or only weakly thermally metamorphosed as shown by the presence of pyrophyllite and andalusite (Ghosh, 1985). Minerals such as chlorite, muscovite, epidote, chloritoid, and recrystallized sheared quartz in the Los Caribes Formation are representative of the greenschist facies. Also, in contrast to rocks of the Roraima Group, folds in conglomerate, sandstone, and phyllitic shale of the Los Caribes Formation are isoclinal or chevron in shape, and the rocks are foliated and have a fracture cleavage at a high angle to bedding (Benaim, 1972; Ghosh, 1985; Lira and others, 1985). Some conglomeratic units are as thick as 60 m; however, the total thickness of this formation has not been determined. The Los Caribes Formation probably represents environments transitional from marine to continental.

Rocks of the Los Caribes Formation are chronologically correlative with those of the Cinaruco Formation in Estado Amazonas and with those of the Muruwa, Rosebel, and Orapu Formations in Guyana, Suriname, and French Guiana, respectively (Ghosh, 1985). These rocks also resemble a gold-bearing sequence known as the Tarkwaian Series in the eastern West African craton (Black, 1980; Bonhomme and Bertrand-Sarfati, 1982; Kesse, 1985; Cohen and Gibbs, 1989; Vinchon, 1989; Ledru and others, 1994). The Tarkwaian Series consists of Early Proterozoic clastic metasedimentary rocks about 2,500–6,700 m thick that are folded within the Birimian greenstone-belt rocks (Ntiamoah-Agyakwa, 1979; Black, 1980; Kesse, 1985; Norman and Appiah, 1989; Eisenlohr and Hirdes, 1992; Watkins and others, 1993).

REAL CORONA–EL TORNO ASSEMBLAGE

Supracrustal rocks, including basal feldspathic quartzite and conglomerate, tholeiitic basalt, gabbro, and thin beds of shale, chert, and ferruginous quartzite, form an east-trending structural basin about 100 km southwest of Ciudad Bolívar (fig. 2). The basin, or syncline, is about 45 km long and 16 km wide and is underlain both on the north and south by a gneissic basement about 2,240 Ma in age (Kalliokoski, 1965; Sidder, Day, and others, 1991). The basal quartzite is in depositional contact with the basement of granitic gneiss. The quartzite is as thick as 150 m; thicknesses of the other units are not known. These rocks have been penetratively deformed; they exhibit well-developed foliation, mineral lineation, and mylonitic fabric and are metamorphosed to the amphibolite facies. Hills underlain by quartzite and chert are as much as 200 m above the generally flat gneissic terrain. Vegetation in the area is savannalike.

AGE, CHEMISTRY, AND ORIGIN

Greenstone belts of the Guayana Shield formed during the Early Proterozoic and closely resemble in structure, lithostratigraphy, and composition of their metavolcanic and metasedimentary rocks and in areal extent greenstone belts of Early Proterozoic age such as those associated with the West African Shield and the Penokean orogen in the Superior province of the Canadian Shield, as well as many greenstone belts of Archean age (Sims and others, 1989; Sylvester and Attoh, 1992). Geochronological data, including uranium-lead zircon and whole-rock samarium-neodymium and rubidium-strontium isochron dates, document that the metavolcanic greenstone-belt rocks and associated granitic rocks were emplaced throughout the Guayana Shield between about 2,250 and 2,100 Ma (Gibbs and Olszewski, 1982; Gruau and others, 1985). These ages are the same as those obtained from the most detailed geochronological study of rocks of the greenstone belts of Venezuela, which determined a conservative age range of less than 2,300 to about 2,050 Ma, and a narrower range of 2,165 to 2,080 Ma, for emplacement of the metavolcanic-metasedimentary greenstone-belt sequence and crystallization of granite of the Supamo Complex (Klar, 1979). A sample of dacitic tuff from the Yuruari Formation in the Lo Increíble mining district has a uranium-lead zircon age of 2,131±10 Ma (Day and others, this volume), which coincides with the published ages of rocks of the greenstone belts. Voluminous magmatism in the Birimian greenstone belts of the Reguibat and Leo Shields of the West African craton has been dated by uranium-lead zircon and samarium-neodymium and lead-lead whole-rock dating methods as about 2.18–2.07 Ga; the maximum interval of 2.3–2.0 Ga includes scattered, less well constrained dates (Cahen and others, 1984; Boher and others, 1992; Hirdes and others, 1992; Taylor and others, 1992).

The chemistry of rocks of the greenstone belts in the Guayana Shield has not been systematically studied. Those investigators who have conducted geochemical studies on these rocks noted that the original chemical composition of the igneous rocks has been altered by weathering, hydrothermal alteration (spilitization and potassium metasomatism), and greenschist- and amphibolite-facies regional metamorphism. Recent studies by Gibbs (1987) and Renner and Gibbs (1987) in Guyana and by Day and others (1989) in Venezuela provide data for representative areas of the Guayana Shield. Both tholeiitic and calc-alkaline differentiation trends are common in the volcanic rocks. Low-potassium subalkaline tholeiitic basalt and basaltic andesite are the dominant rock types; lesser subequal proportions of calc-alkaline andesite, dacite, and rhyolite are also present (Gibbs, 1987). Few rocks in the lower part of the greenstone belts from throughout the Guayana Shield have silica contents in the range from 63 to 68 weight percent, and the overall distribution of silica may be considered bimodal (Gibbs, 1987); however, rocks of the Caballape Formation in the Anacoco area of Venezuela have a compositional continuum with a systematic variation in the major and trace elements that forms a cogenetic mafic to felsic calc-alkaline magmatic series (Day and others, 1989). The Birimian greenstone belts of the West African craton also include tholeiitic basalt overlain by calc-alkaline intermediate to felsic tuff and lava (Boher and others, 1992; Sylvester and Attoh, 1992; Vidal and Alric, 1994).

Early Proterozoic magmatism in the Guayana Shield (and in its associated West African craton) represents a major period of rapid crustal growth from mantle-derived melts; however, discriminant diagrams used to define the tectonic setting of volcanic rocks are not consistent in specifying the tectonic environment of deposition of the greenstone-belt rocks of the Guayana Shield (Gibbs, 1987). The volcanic rocks were erupted predominantly in a submarine environment, and they have chemical characteristics of modern ocean floor basalts, island arc rocks, and continental arc rocks. Initial $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios of 0.7019 and 0.51002, respectively, as well as an ε_{Nd} value of 2.1 (Gruau and others, 1985), indicate that the volcanic rocks are derivatives of mantle melts and do not contain any component of Archean continental crust. Similarly, the Birimian volcanic and granitic rocks contain a negligible component, if any, of Archean crust (Rocci and others, 1991; Boher and others, 1992; Sylvester and Attoh, 1992; Taylor and others, 1992). Ambiguous geochemical signatures in the volcanic rocks of the Guayana Shield (and West African craton) may reflect an evolutionary history from young, immature, oceanic island arc to back-arc marginal basin and mature (thickened) island arc environments. The Trans-Amazonian orogeny (Eburnean orogeny in West Africa) resulted in the accretion of these new crustal components (Choudhuri, 1980; Bosma and others, 1983; Gibbs, 1987; Renner and Gibbs, 1987; Teixeira and others, 1989; Liégeois and others, 1991; Rocci and others, 1991; Boher and others, 1992).

Supracrustal rocks of the Real Corona–El Torno assemblage were deposited in a submarine environment within or marginal to the craton, possibly as a back-arc marginal basin inboard from the Pastora island arc or during an early cratonic rifting phase associated with the Trans-Amazonian orogeny. Closure of the basin and regional compressive tectonism associated with the later phases of the Trans-Amazonian orogeny produced penetrative deformation and metamorphism and variably thrust the basaltic rocks over the basal quartzite. The age of the basement gneiss places a maximum age on the rifting of 2,240 Ma (Sidder, Day, and others, 1991).

MINERAL DEPOSITS

Rocks of the greenstone belts of the Guayana Shield contain deposits of shear-zone-hosted, low-sulfide gold-quartz veins (Berger, 1986). Several deposits in Venezuela are currently being mined, including those in the El Callao, Lo Increíble, and Botanamo districts (Sidder, this volume). Mafic metavolcanic rocks of the El Callao Formation are the most common ore host in the El Callao district, which is the largest gold-producing district in Venezuela and has total production of about 200 metric tons of gold; however, all rock types in the greenstone belts throughout the shield, except those in the Los Caribes Formation, are known to host low-sulfide gold-quartz veins (Korol, 1961; Carter and Fernandes, 1969; Menendez, 1972; Dahlberg, 1975; Blanc and others, 1980; Barnard, 1990). Ore in the greenstone belts is localized by faults and shear zones (Menendez, 1972; Day and others, this volume). Quartz veins range in thickness from 2 cm to more than 10 m. Quartz is milky white to gray and locally banded. Native gold, minor to trace amounts of pyrite, and lesser amounts of tetrahedrite, chalcopyrite, bornite, molybdenite, scheelite, and sphalerite are the most typical metallic minerals in the quartz veins. Carbonate (commonly ankerite) in the quartz veins and carbonate alteration as much as 30 m into the wallrocks are common in some districts such as El Callao. In addition to carbonate alteration, the wallrocks are intensely silicified, sericitized, and propylitized (with epidote and chlorite) as much as several tens of meters from the veins, and tourmaline and mariposite (chrome mica) are variably present in the alteration assemblage (Macdonald, 1968; Banerjee and Moorhead, 1970; Barron, 1973; Menendez, 1974). These shear-zone-hosted, low-sulfide gold-quartz vein deposits are similar to Early Proterozoic gold deposits in Ghana and elsewhere in West Africa (Milési and others, 1992; Dzigbodi-Adjimah, 1993; Petersen, 1993). The quartz-vein-hosted and other gold deposits in Early Proterozoic rocks of the greenstone belts of West Africa, excluding the Tarkwaian gold-bearing conglomerates, have produced more than 1,000 metric tons of gold at grades between 2 and 30 grams per metric ton, and reserves of more than 350 metric tons are present (Milési and others, 1989, 1992; Dzigbodi-Adjimah, 1993).

Vuelvan Caras and Payapal are the only gold-bearing vein deposits in the Venezuelan Guayana Shield known to be present in granitic rocks (Graterol, 1974; Wynn and Sidder, 1991). They are similar to gold deposits in Guyana, such as those at Omai, Peters' Mine, and Eagle Mountain, that are in stockwork veins within and at the periphery of granitic plutons that range in age from about 2,100 to 1,800 Ma as dated by whole-rock rubidium-strontium isochrons (Macdonald, 1968; Barron, 1969; Carter and Fernandes, 1969; Berrange, 1977; Elliot, 1986; Barnard, 1990; Gibbs and Barron, 1993). Although in Venezuela the association of granitic intrusive rocks and gold-bearing quartz veins is not well documented, several authors have noted that gold deposits in greenstone-belt rocks in Guyana, Suriname, and French Guiana are spatially associated with, if not genetically related to, granitic intrusive rocks (Carter and Fernandes, 1969; Banerjee and Moorhead, 1970; Dahlberg, 1975, Blanc and others, 1980; Gibbs and Barron, 1993). Some gold deposits within the Birimian greenstone belts of Ghana in West Africa show a spatial association with granitic intrusive rocks (Leube and others, 1990). Early Proterozoic granitic rocks associated with the Birimian greenstone belts throughout West Africa contain gold reserves of less than 50 metric tons, which is a minor proportion (about 3 percent) of the total gold reserves in Early Proterozoic rocks (Milési and others, 1992).

Metals other than gold have not been produced in any significant quantity from rocks of the greenstone belts of the Guayana Shield. Small prospects of manganese in Venezuela, such as San Cristobal and La Esperanza, and iron prospects in banded iron formation have not been systematically worked or evaluated. The manganese prospects in Venezuela are similar to sedimentary-volcanogenic deposits at Matthews Ridge in Guyana, Serra do Navio in Amapa Federal Territory, Brazil, and Nsuta in Ghana, Tambao in Burkina Faso, and Ziémougoula, Ivory Coast, of West Africa. Matthews Ridge produced about 1.3 million metric tons with a grade of about 39.3 percent manganese from 1961 to 1968 and has ore reserves of about 288,500 metric tons of 37 percent manganese (Barron, 1973). Prospects in the San Cristobal area may be extensions of manganiferous strata from the Matthews Ridge area (Gibbs and Barron, 1993). Serro do Navio has measured reserves of 15.7 million metric tons at 39.3 percent MnO_2, and total reserves of 20.1 million metric tons (Nagell, 1962; Damasceno, 1982; Lima, 1984; Schobbenhaus and others, 1984). Production from Nsuta between 1954 and 1983 totalled about 12.8 million metric tons of manganese (Kesse, 1985). Reserves are estimated to be about 5 million metric tons of high-grade oxide ore (48.9 percent manganese) and 2 million metric tons of low-grade oxide ore (34.3 percent manganese) and about 28 million metric tons of carbonate ore (26.9 percent manganese) (Kesse, 1985). Some of the manganese horizons in Ghana have a spatial relation with gold deposits (Leube and others, 1990; Milési and others, 1992). The manganese deposits are on the flanks of the volcanic belts and in the transition zone from volcanic rocks to basinal sedimentary rocks. These transition zones, or breaks, are characterized by chemical sedimentary rocks such as chert, iron-magnesium carbonate rocks, and carbon-rich argillite and are the foci for the gold deposits (Leube and others, 1990; Milési and others, 1992).

Undiscovered deposits such as those of Algoma-type banded iron formation and Homestake-type gold may be present in the Early Proterozoic metavolcanic and metasedimentary rocks of the greenstone belts of the Venezuelan Guayana Shield. Banded iron formation is generally absent in the Guayana and West African Shields; however, manganese-rich exhalative rocks are common (Holtrop, 1965; Leube and others, 1990; Milési and others, 1992). As discussed previously, manganese-rich rocks and other chemical sedimentary rocks are important exploration guides for gold deposits in the Early Proterozoic rocks of the greenstone belts in Ghana (Ntiamoah-Agyakwa, 1979; Leube and others, 1990).

Volcanogenic massive sulfide deposits have not been discovered in the Guayana Shield, although the tectonic and volcanosedimentary environments are favorable for volcanic-hosted kuroko-type massive sulfide deposits. Prospects in Guyana such as Groete Creek, near Aremu, and in the middle Puruni River area contain anomalous copper, zinc, and gold in metavolcanic and metasedimentary sequences (Barron, 1973; Gibbs and Barron, 1993). Polymetallic volcanogenic massive sulfide deposits are also uncommon in the West African craton (Milési and others, 1992). The only known Early Proterozoic volcanogenic massive sulfide deposit in West Africa is Perkoa in Burkina Faso. It is a zinc-silver deposit dated at 2,120±41 Ma (Marcoux and others, 1988); reserves are estimated to be 4.5 million metric tons at 17 percent zinc and about 60 grams silver per metric ton (Milési and others, 1989, 1992). Although bands or layers of sulfide-rich rocks are rare in the West African Shield, disseminated pyrite and arsenopyrite are common in gold-bearing carbonate within the sequence of chemical sedimentary rocks in the transition zone from volcanic rocks to basinal sedimentary rocks (Leube and others, 1990).

Platinum-group elements, chromium, and nickel-copper deposits may be associated with mafic-ultramafic complexes in the greenstone belts. Anomalous amounts of platinum-group elements are present in mafic rocks at Pistón de Uroy and in the Real Corona–El Torno assemblage (Sidder, this volume; Wynn, Page, and others, this volume). Platinum-group elements are also present in gold lode and placer deposits (Dahlberg, 1975; Sidder, this volume).

SUPAMO COMPLEX

The Supamo Complex includes paragneiss, schist, migmatite, and granitic rocks such as trondhjemite, tonalite, and granodiorite (Moreno and Mendoza, 1975). Quartz monzonite and granite that intrude trondhjemite, tonalite, and granodiorite and the greenstone-belt rocks of the Pastora Supergroup have also been included within the Supamo Complex by some authors (Menendez, 1972). Plutonic rocks of the Supamo Complex are massive to foliated and generally form domes; metamorphic grade in the country rocks increases from greenschist to amphibolite facies within about 6 km of the intrusive bodies. The marginal facies of the intrusive bodies are generally concordant with the supracrustal host rocks. The granitic rocks generally underlie a savanna that has small, rounded, isolated hills and a dendritic drainage pattern. The soil is sandy with minor clay and has a bleached or whitish color (Menendez, 1972, 1974; Benaim, 1974). Rocks of the Supamo Complex are not known to host any mineral deposits.

The age of the Supamo Complex is commonly described as ranging from 2,700 to 2,100 Ma; the younger rocks are said to be "remobilized Supamo." Recent uranium-lead isotopic data for zircons indicate that trondhjemite of the Supamo Complex crystallized between about 2,200 and 2,050 Ma (Klar, 1979). The age of emplacement of quartz monzonite plutons into the greenstone-granite terrane and the Imataca Complex also ranges from about 2,200 to 2,050 Ma, on the basis of uranium-lead zircon dates (Klar, 1979), and possibly to 1,958 Ma, as indicated by rubidium-strontium data for biotite (Onstott, Hargraves, York, and Hall, 1984). The reinterpreted age of gneissic rocks in the Supamo Complex is about 2,230 Ma, consistent with an age of 2,227 Ma for the apparently correlative Bartica Gneiss in Guyana (Gibbs and Olszewski, 1982). The maximum age for gneiss in the Supamo Complex is about 2,300 Ma (Klar, 1979). Granite and high-grade gneiss associated with the greenstone belts of the West African craton also range in age from about 2,220 to about 2,070 Ma (Rocci and others, 1991; Boher and others, 1992; Taylor and others, 1992). Trondhjemitic intrusive rocks and more potassium rich granite and granodiorite of the West African craton have similar ages (Boher and others, 1992), as is true in the Venezuelan Guayana Shield. Sodic granitoid rocks associated with the Nangodi greenstone belt in Ghana have high Sr/Nd and Tb/Dy ratios relative to normal mid-ocean ridge basalt that indicate the granitic rocks may have formed by melting of relatively hot, subducted slab material ("amphibolite") (Sylvester and others, 1993). Essentially contemporaneous potassic-rich granitic rocks, therefore, may have had less of a slab component and incorporated more sialic crustal material. Gneiss and migmatite in Ghana are coeval and cogenetic with the granitic rocks (Opare-Addo and others, 1993). Differences in structural and textural features between the metamorphic and granitic rocks, such as foliation, homogeneity, and grain size, are due to emplacement at different depths; the migmatites were generated at deeper crustal levels than the granitic rocks (Opare-Addo and others, 1993).

TRANS-AMAZONIAN OROGENY

The Trans-Amazonian orogeny was a major cycle of greenschist- to upper amphibolite- and granulite-facies metamorphism, deformation, and magmatic activity in the Guayana Shield during the Early Proterozoic. It was a period of continental collision and accretion of assorted Archean and Early Proterozoic terranes into the Amazonian craton, the subsequent common deformation of these accreted terranes, and the first development of continental environments on much of the craton. Paleomagnetic and $^{40}Ar/^{39}Ar$ data indicate that the Guayana and Guaporé Shields and possibly the West African craton were combined as a single tectonic plate during the Trans-Amazonian orogeny (Gibbs and Wirth, 1986; Renne and others, 1988; Teixeira and others, 1989; Rocci and others, 1991; Boher and others, 1992). The Imataca Complex was thrust over the Pastora-Supamo greenstone-granite terrane during the Trans-Amazonian orogeny, and oblique compression between these two terranes resulted in left slip along major fault zones such as the Guri shear zone (Swapp and Onstott, 1989). The Guri fault zone, and its inferred continuation in the West African craton, the Sassandra–Trou Mountain fault zone (Caen-Vachette, 1988; Cohen and Gibbs, 1989), is a suture between Archean rocks (the Imataca Complex in Venezuela) and Early Proterozoic greenstone-belt rocks.

The Trans-Amazonian orogeny in the Guayana Shield occurred between about 2,150 and 1,960 Ma and possibly continued to about 1,730 Ma. The wide range in ages may be the result of tectonic activity that was time transgressive across the shield from northeast to southwest (Gaudette and Olszewski, 1985). Alternatively, the range in ages may span two distinct orogenic episodes, a predominantly collisional and metamorphic event accompanied by intrusive magmatic activity throughout the shield from about 2,150 to 1,960 Ma, and a period of intense metamorphism, deformation, and intrusion between about 1,860 and 1,730 Ma in the westernmost part of the shield in Estado Amazonas of Venezuela (Klar, 1979; Gibbs, 1980; Bosma and others, 1983; Teixeira and others, 1984; Gaudette and Olszewski, 1985). Postorogenic and (or) anorogenic magmatic activity from about 1,930 to about 1,790 Ma that emplaced rocks of the Cuchivero Group and its equivalents and was accompanied by minor uplift but little or no deformation or metamorphism is not considered here to be part of the Trans-Amazonian orogeny.

Rocks of the Imataca Complex reveal a history of prograde metamorphism and retrograde cooling and uplift associated with the Trans-Amazonian orogeny. Detailed rubidium-strontium isotopic studies of the age of granulite-facies metamorphism in the Imataca Complex indicate an age of 2,022±67 Ma for the Trans-Amazonian orogeny (Montgomery and Hurley, 1978). This age apparently predates the peak pressure associated with the orogeny and postdates the peak temperature (Swapp and Onstott, 1989). Plateau dates from $^{40}Ar/^{39}Ar$ analyses of hornblende and biotite in the Imataca Complex range from about 2,044 to 1,760 Ma and are interpreted to record decompression, uplift, and cooling following peak metamorphism of the Trans-Amazonian orogeny (Onstott and others, 1989; Swapp and Onstott, 1989). The Imataca Complex was uplifted from a depth of about 32 km to about 16 km during southward thrusting of the Imataca terrane over the greenstone belts within about 30 m.y. after initiation of uplift and after reaching peak pressure conditions (Swapp and Onstott, 1989). Extremely slow cooling rates implied by plateau dates of biotite, potassium feldspar, and plagioclase from granulite-facies rocks in the core of the Imataca Complex indicate that uplift of the Imataca Complex had mostly ceased by about 1,962 Ma and that rocks in the complex cooled isobarically at intermediate crustal levels of about 15 km until about 1,100 Ma (Onstott and others, 1989; Swapp and Onstott, 1989). All argon-bearing mineral systems were closed by 1,100 Ma, perhaps as a result of renewed uplift associated with the Nickerie or K'Mudku episode (table 1) (Onstott and others, 1989). In contrast, metamorphic rocks in the Carajas region, Brazil, of the Guaporé Shield cooled rapidly and attained stable magnetization by about 1,910 Ma (Renne and others, 1988). Paleomagnetic and $^{40}Ar/^{39}Ar$ data preclude the possibility of any regional metamorphism in the Guaporé Shield to even the lowest greenschist facies after the Trans-Amazonian orogeny (Renne and others, 1988).

The Trans-Amazonian orogeny in the Guayana Shield is defined here to consist of deformation, metamorphism, and magmatic activity between about 2,150 and 1,960 Ma. Rocks of the Pastora Supergroup were deformed in two pulses or episodes of tectonic activity within this interval, whereas rocks of the Botanamo Group were affected only by the second pulse of deformation.

The Eburnean orogeny in the West African craton was a period of igneous and metamorphic activity and deformation about 2,200–1,980 Ma, possibly between only 2,112 and 2,073 Ma (Hedge and others, 1975; Onstott and Dorbor, 1987; Cohen and Gibbs, 1989; Feybesse and others, 1989; Liégeois and others, 1991), coincident with the Trans-Amazonian orogeny in the Guayana Shield. Deformation associated with the Eburnean orogeny in the West African craton has been subdivided into two phases; the lower group of Birimian rocks (Birimian I) was affected by both phases, and the upper group (Birimian II) was folded just once (Rocci and others, 1991; Milési and others, 1992). Recent work suggests, however, that the distribution of deformation features is spatially controlled by shear zones and is not temporally restricted to particular stratigraphic units (Mortimer, 1992; Watkins and others, 1993). Hence, all Birimian and Tarkwaian units have been affected by all phases of the Eburnean orogeny (Mortimer, 1992: Watkins and others, 1993). Deformation D1 is represented by thrust faults (Feybesse and Milési, 1994), foliation S1 that is subparallel with bedding (S0), lineation L1 defined by the intersection of bedding and foliation S1, isoclinal folds P1, and low- to medium-grade metamorphism; it is referred to as a tangential (S1 parallel to S0) deformation event (Feybesse and others, 1989; Milési and others, 1989, 1992; Liégeois and others, 1991; Eisenlohr and Hirdes, 1992). Deformation D1 is attributed to collision tectonics. Calc-alkaline volcanism and deposition of volcaniclastic and turbiditic material in the upper Birimian, as well as deposition of the Tarkwaian sediments (equivalent to the Caballape and Los Caribes Formations in the El Callao area and the so-called eugeosynclinal metavolcanic rocks in the Botanamo and Anacoco areas [Cox, Wynn, and others, 1993]), followed deformation D1. Deformation D2 was a sinistral transcurrent shear event characterized by left-lateral strike-slip faults that are locally associated with thrust zones, large, regional upright folds P2, a crenulation cleavage S2 that crenulates foliation S1, stretching lineation L2, and low- to medium-grade metamorphism (Feybesse and others, 1989; Liégeois and others, 1991; Eisenlohr and Hirdes, 1992; Milési and others, 1992; Vidal and Alric, 1994). This second deformation was also a period of compressive transcurrent tectonics. Uranium-lead, samarium-neodymium, and rubidium-strontium dates determined for D1 and D2 in West Africa indicate that these deformations associated with the Eburnean orogeny took place between about 2,112 and 2,073 Ma (Feybesse and others, 1989; Liégeois and others, 1991). In the eastern part of the West African shield, deformation D3 produced dextral shear zones, folds P3, foliation and crenulation cleavage S3, a stretching lineation L3, and reactivation and local offset of D2 faults (Milési and others, 1989, 1992; Liégeois and others, 1991; Watkins and others, 1993). Event D3 accompanied low-grade metamorphism at about 2,073 Ma (Milési and others, 1989). A late or post-Eburnean tectonic shortening event characterized by a crenulation cleavage and a transcurrent shear zone, but unaccompanied by igneous or metamorphic activity, may have occurred as many as 90 m.y. after deformation D3 at about 1,980 Ma (Milési and others, 1989; Liégeois and others, 1991; Vidal and Alric, 1994).

The Cuchivero Group, about 1,930–1,790 Ma, is regarded here to be post-Trans-Amazonian in age. Postcollisional magmatism following collision and amalgamation of the island-arc greenstone belts with the Imataca Complex resulted in the eruption and intrusion of the Cuchivero Group and its equivalents. Rocks of the Cuchivero Group are generally unmetamorphosed and relatively undeformed, and their character is sufficiently different from that of rocks of the greenstone belts that they should be considered part of an unnamed thermomagmatic event. This post-Trans-Amazonian, postcollisional magmatic activity continued until about 1,790 Ma. Similar, but less voluminous, posttectonic magmatism in the Reguibat Shield of the West African craton took place between about 1,970 and 1,750 Ma (Cahen and others, 1984; Rocci and others, 1991).

Metamorphism, deformation, and magmatic activity about 1,860–1,730 Ma in Estado Amazonas must be considered as part of a separate orogenic event (see discussion of undivided Proterozoic rocks). Gaudette and Olszewski (1985) correlated this younger orogenic episode in Estado Amazonas with the older Trans-Amazonian event (2,150–1,960 Ma); however, the age of 1,900±200 Ma reported for the Trans-Amazonian orogeny (Moreno and others, 1977; Gaudette and Olszewski, 1985), which is based on potassium-argon and whole-rock rubidium-strontium dates for an intrusion in the Imataca Complex and for rocks of the Cuchivero Group, is not compatible with the recent and more accurate uranium-lead and argon-argon dates for metamorphism and deformation of the Trans-Amazonian orogeny. As noted in the discussion on undivided Proterozoic rocks in Estado Amazonas, geologic mapping, geochemical sampling, and geochronologic dating are insufficient for a rigorous interpretation of the geology in this area.

CUCHIVERO GROUP

Early to Middle Proterozoic supracrustal rocks were emplaced in and deposited on the older greenstone-granite terrane in the southern, central, and western parts of the Venezuelan Guayana Shield (fig. 2). They include a thick pile (greater than 3 km thick) of mostly felsic to lesser intermediate and mafic volcanic, subvolcanic, and plutonic rocks and associated volcanogenic sedimentary rocks of the Cuchivero Group; sandstone and conglomerate and lesser siltstone, shale, chert, and interlayered felsic volcanic rocks (jasper) of the Roraima Group; continental tholeiitic dikes, sills, inclined sheets, and small irregular intrusive bodies of the Avanavero Suite; and rapakivi granite of the Parguaza Granite (Rios, 1972; Mendoza and others, 1975; Ghosh, 1985; Gibbs, 1986; Sidder and Martinez, 1990).

The Cuchivero Group includes the relatively older volcanic rocks of the Caicara Formation and the younger granites of Guaniamito, San Pedro, and Santa Rosalía (Rios, 1972; Mendoza and others, 1975). Mesozoic and (or) Cenozoic tilting of the shield and subsequent erosion have exposed the deeper level granitic rocks in the northern part of the Cuchivero terrane, whereas high-level volcanic rocks of the Caicara Formation predominate in the southern part. Broad, open folds are common, and structures such as faults and lineaments generally strike northwest to north-northwest and north-northeast. The El Viejo Formation, the volcanic suite of the Parucito Valley, and other relatively unmetamorphosed volcanic and plutonic rocks in Estado Amazonas, the Uatumã Supergroup (including the Surumu and Iricoumé Formations and the granodiorite of Serra do Mel) in northern Brazil, the Kuyuwini and Burro-burro Groups in Guyana, and the Dalbana Formation in Suriname all correlate with the Cuchivero Group (Mendoza and others, 1975; Montalvao, 1975; Talukdar and Colvée, 1975, 1977; Berrange, 1977; Mendoza and others, 1977; Tepedino, 1985; Gibbs, 1987; Sidder and Martinez, 1990; Machado and others, 1991; Gibbs and Barron, 1993; Dall'Agnol and others, 1994). Minor amounts of rocks equivalent to the Cuchivero Group are present in French Guiana and the West African craton (Cahen and others, 1984; Gibbs, 1987; Rocci and others, 1991).

CAICARA FORMATION

The Caicara Formation consists of subaerially deposited pyroclastic rocks including variably welded ash-flow and air-fall tuff and breccia and minor lava flows and domes and intercalated volcaniclastic rocks. The rocks are aphyric to porphyritic, and both crystal-rich and crystal- and lithic-rich varieties are present. Vitroclastic and eutaxitic textures including devitrified glass shards and collapsed pumice fragments are common (Rios, 1972; Mendoza, 1977b; Sidder and Martinez, 1990). The Caicara Formation is made up of rhyolite, subordinate rhyodacite and dacite, and minor proportions of andesite, basaltic andesite, and basalt. On a total alkali-silica diagram, some rocks of the Caicara Formation are classified as trachyte and trachydacite. Silicic rocks are mainly tuff, whereas andesitic and basaltic rocks are lava flows and dikes. These rocks are metaluminous to peraluminous, subalkalic to alkalic, and together form an apparent comagmatic calc-alkaline series (Talukdar and Colvée, 1975, 1977; Mendoza, 1977b; Sidder and Martinez, 1990). Felsic to intermediate volcanic and volcaniclastic rocks near Ichún Tepuy (fig. 1) named the Ichún Formation are interpreted to be part of the Roraima Group (Briceño and others, 1989); however, these rocks are chemically and lithologically similar to rocks of the Caicara Formation to the north and south along the Río Paragua and have the same stratigraphic position. Moreover, the basal conglomerate and overlying sandstone of the Roraima Group near Ichún Tepuy conformably overlie the volcanic rocks (Sidder, unpublished data, 1988). A similar conformable contact has been mapped between rocks of the Caicara Formation and the overlying Roraima Group in Estado Amazonas (Stephen D. Olmore, U.S. Geological Survey, oral commun., 1990), and intercalations of volcanic tuff of the Surumu Formation are common near the base of the Roraima Group in Brazil within 80 km south of the Brazil-Venezuela-Guyana junction (Amaral and Halpern, 1975).

GRANITE OF THE CUCHIVERO GROUP

Granite associated spatially and temporally with the volcanic rocks of the Caicara Formation includes the granites of San Pedro, Santa Rosalía (including the granite of Las Trincheras), and Guaniamito. These granitic bodies comprise hypabyssal biotite granite, quartz monzonite, and granodiorite (Rios, 1972; Mendoza, 1974; Tepedino, 1985) and are in intrusive and fault contact with the volcanic rocks. The rocks are generally equigranular to porphyritic and medium to coarse grained. The granite of San Pedro is dominantly a fine-grained leucocratic granite that has been interpreted as a marginal border phase of the coarser grained biotite granite of Santa Rosalía (Mendoza, 1974). The granite of Guaniamito is a porphyritic, medium to coarsely crystalline, biotite±hornblende granite. All of these granites are generally massive in texture but locally are foliated, especially near the intrusive contact of granite with the volcanic rocks of the Caicara Formation (Rios, 1972). Primary minerals include potassium feldspar (orthoclase and microcline, 20–60 modal percent), quartz (10–40 percent), plagioclase (albite-oligoclase, 5–40 percent), biotite (<1–10 percent), and accessory sphene, apatite, zircon, muscovite, hornblende, allanite, and iron-titanium-oxide minerals (magnetite and lesser ilmenite). Secondary alteration minerals are epidote, clinozoisite, and white mica in plagioclase and potassium feldspar, chlorite after biotite, and hematite after magnetite. Aplite dikes and barren quartz veins commonly cut the granitic bodies (Rios, 1972; Mendoza, 1974; Tepedino, 1985; Sidder, unpublished data, 1988). Hypabyssal biotite granite and leucogranite in Suriname that are associated with and intrude rhyolitic volcanic rocks of the Dalbana Formation (De Roever and Bosma, 1975; Bosma and others, 1983) are considered to be comagmatic equivalents of the volcanic rocks (Mendoza, 1977b; Bosma and others, 1983).

Granitic and volcanic rocks of the Cuchivero Group and their equivalents throughout the Guayana Shield are generally unmetamorphosed. Reports of lower greenschist facies metamorphism apparently refer to local contact metamorphic aureoles in volcanic and volcaniclastic rocks close to intrusions. For example, field and petrographic evidence of metamorphism is not present in samples of volcanic rocks from the upper Río Caura or Río Paragua areas (Sidder, unpublished data, 1990; Sidder and Martinez, 1990). Indeed, as noted previously, vitroclastic and eutaxitic textures are abundant. Minor alteration in the volcanic and granitic rocks, such as partial replacement of feldspar by fine-grained sericite and epidote, chlorite alteration of biotite, and thin veinlets of quartz or epidote+chlorite, is indicative of deuteric and local hydrothermal alteration. Rios (1972) recognized that the volcanic rocks of the Caicara Formation are thermally metamorphosed in restricted zones close to intrusions. It is significant to note that Rios called primary flow bands developed during the extrusion and emplacement of the volcanic rocks "foliación," or foliation. He did not recognize any effects of regional metamorphism. Other authors have, however, referred to the Caicara Formation as a sequence of metavolcanic rocks (Tepedino, 1985), and still others have called the rocks metavolcanic but have noted that the rocks have undergone contact metamorphism ("thermal-regional" or "plutono-metamorphism") only due to intrusion of the granitic batholith of the Cuchivero Group (Mendoza, 1977b). Similarly, Bosma and others (1983) referred to rhyolitic rocks of the Dalbana Formation in Suriname as metavolcanic but noted (p. 247) that "Distinctly recrystallized metavolcanics***form broad marginal zones along the granites***The recrystallization, without significant foliation or folding, is spatially related to granite intrusions and probably took place at shallow depth under hornblende-hornfels facies conditions." Thus, the volcanic rocks of the Caicara Formation and granitic rocks of the Cuchivero Group and similar rocks throughout the Guayana Shield are not regionally metamorphosed, but they are contact metamorphosed or hydrothermally altered in proximity to intrusions, faults, dikes, and veins (De Roever and Bosma, 1975; Sidder and Martinez, 1990).

AGE AND ORIGIN

Whole-rock rubidium-strontium dates for rocks of the Cuchivero Group and its equivalents throughout the Guayana Shield range from 1,930 to 1,640 Ma, and the majority are between 1,930 and 1,790 Ma (table 2) (Hurley and others, 1977; Moreno and others, 1977; Bosma and others, 1983; Schobbenhaus and others, 1984; Teixeira and others, 1989; Machado and others, 1991). Initial $^{87}Sr/^{86}Sr$ ratios range from 0.698 for the granites of Guaniamito, San Pedro, and Santa Rosalía to 0.721 for volcanic rocks of the Surumu Formation in Brazil (table 2). The former initial ratio is not geologically reasonable because basaltic achondrite meteorites have an initial ratio of 0.69897 (Faure, 1986). The latter ratio is extremely high due to scatter of the data; the mean squares weighted deviation (MSWD) is 60.9, and the 1-sigma error in the calculated initial ratio is 0.017. Most of the initial ratios for rocks of the Cuchivero Group and its equivalents are between about 0.705 and 0.707, values between initial ratios inferred for Proterozoic crust (>0.708) and mantle-derived magmas (<0.7045) (Priem, 1987; Riciputi and others, 1990), and are a further indication of some contribution of continental crust in the generation of the magmas. Those initial ratios as high as 0.712 may reflect melts derived directly from crustal material.

Rocks of the Cuchivero Group and their equivalents have been considered both as orogenic, related to the Trans-Amazonian collision and deformation, and as anorogenic (De Vletter and Kroonenberg, 1984). Supporters of the orogenic interpretation suggest that the contact between the felsic volcanic rocks and rocks of the underlying greenstone belt is conformable, and therefore they include the Cuchivero Group and its equivalents as a second, dominantly magmatic, stage of the Trans-Amazonian orogeny between about 2,000 and 1,870 Ma (Bosma and others, 1983; De Vletter and Kroonenberg, 1984; Teixeira and others, 1984). As noted above, however, characteristics of deformation and metamorphism in the Cuchivero Group are significantly different from those in rocks of the greenstone belts, and therefore a conformable contact is extremely unlikely. Those who suggest that the predominantly felsic igneous rocks of the Cuchivero Group and its equivalents are anorogenic have noted that the contact with rocks of the greenstone belts is a profound angular unconformity and that crustal extension, as evidenced by widespread intrusion of mafic dikes and sills of the Avanavero Suite, was penecontemporaneous with deposition of the volcanic rocks (Montalvao, 1975; Gibbs, 1980, 1986; Gibbs and Olszewski, 1982). As discussed following, however, the Avanavero Suite (about 1,650 Ma) is younger than the Cuchivero Group, and these units are not coeval.

The volcanic and plutonic rocks of the Cuchivero Group are referred to here as postcollisional, post-Trans-Amazonian because they do not have a clear association with any orogenic belt, they are not regionally metamorphosed, and they are weakly deformed and lack a pervasive penetrative fabric. Any deformation that they underwent may be attributed to younger post-Trans-Amazonian events, and low-grade contact metamorphism in the volcanic rocks is related to granitic intrusions. Granites formed in postcollisional tectonic settings commonly postdate the collisional event by about 25–75 m.y. (Sylvester, 1989), which is the approximate amount of time between termination of uplift of the Imataca Complex (collision with rocks of the greenstone belts) and magmatism in the Cuchivero Group. Limited geochemical data suggest that the tectonic environment of the granite of Santa Rosalía and the volcanic rocks of the Caicara Formation is transitional; the

Table 2. Rubidium-strontium whole-rock isochron dates for volcanic and plutonic rocks of the Cuchivero Group and its equivalents, Guayana Shield.

[Number of samples on which age is based is given in parentheses after age. MSWD is mean squares weighted deviation]

Country	Unit	Age (Ma)	$(^{87}Sr/^{86}Sr)_0$	MSWD	Reference
Venezuela	Caicara Formation	1,700±220 (*n*=3)	0.709	1.72	Gaudette and others (1978).
Venezuela	Granites of Santa Rosalía and San Pedro	1,880±88 (*n*=7)	0.698	24.3	Gaudette and others (1978).
Guyana	Kuyuwini Group	1,800±420 (*n*=4)	0.705	20.9	Berrange (1977).
Suriname	Felsic volcanic rocks (Dalbana Formation)	1,930±48 (*n*=18)	0.705	2.25	Priem and others (1971).
Suriname	Granitoid rocks	1,850±40 (*n*=14)	0.707	2.58	Priem and others (1971).
Suriname	Granitic and volcanic rocks together[1]	1,880±31 (*n*=32)	0.706	2.64	Priem and others (1971).
Brazil	Surumu Formation	1,800±94 (*n*=6)	0.721	60.9	Basei and Teixeira (1975).
Brazil	Surumu Formation	1,640±55 (*n*=6)	0.714	18.2	Amaral and Halpern (1975).
Brazil	Surumu Formation[2]	1,820±55 (*n*=10)	0.712	45.7	Basei and Teixeira (1975).
Brazil	Granite of Serra do Mel	1,790±62 (*n*=4)	0.706	4.63	Basei and Teixeira (1975).

[1]As reported by Priem and others(1971).

[2]Includes four samples analyzed by Amaral and Halpern (1975) as reported by Basei and Teixeira (1975).

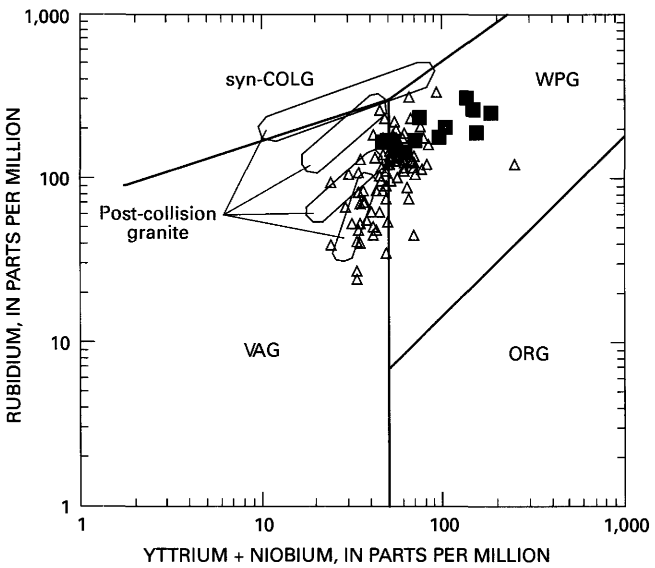

Figure 3. Rubidium versus yttrium+niobium for granite and volcanic rocks of the Early Proterozoic Cuchivero Group of the Guayana Shield. Fields of postcollision granite from Pearce and others (1984); solid squares, granite; open triangles, volcanic rocks; VAG, volcanic-arc granite; ORG, ocean-ridge granite; WPG, within-plate granite; syn-COLG, syncollision granite. Data from Sidder and Martinez (1990) and Sidder (unpublished data, 1990).

granitic and volcanic rocks plot between within-plate (anorogenic) granite and volcanic-arc granite on several discriminant diagrams and approximately in the field of postcollision granite on the rubidium versus yttrium+niobium diagram (fig. 3) (Pearce and others, 1984; Sylvester, 1989; Sidder, unpublished data, 1990). The variously indicated tectonic settings may result from transitional mantle and crustal sources; postcollision, within-plate crustal magmatism may have been followed by renewed subduction and mantle-derived continental arc magmatism. Post-Eburnean granitoid rocks in the West African craton are posttectonic or anorogenic rocks that were intruded at shallow levels and plot in the field of within-plate granite (Rocci and others, 1991).

MINERAL RESOURCES

Major mineral deposits have not yet been discovered in the Cuchivero Group. Quartz veins containing silver and gold are present locally in volcanic rocks of the Caicara Formation, and isolated areas of rhyolite contain trace amounts of disseminated cassiterite (Sidder and Martinez, 1990; Sidder, Brooks, and others, 1991; Sidder, this volume). Although epithermal and bonanza-type precious-metal vein deposits are uncommon in Precambrian rocks (Hutchinson, 1987), the felsic to intermediate composition and pyroclastic character of the volcanic rocks of the Cuchivero Group, as well as geochemical anomalies of silver, bismuth, and molybdenum in some quartz-sulfide veins and gold in panned concentrates, are suggestive of epithermal precious-metal deposits (Sidder and Martinez, 1990; Sidder, Brooks, and others, 1991). Mineralized epithermal systems in volcanic rocks of the Cuchivero Group may have been preserved by burial shortly after deposition by sedimentary rocks of the Roraima Group, as exhibited by locally conformable contacts.

Field and geochemical evidence indicates that shallow porphyritic granitic intrusive rocks in volcanic rocks of the Caicara Formation in western Estado Bolívar and Estado Amazonas and possibly equivalent rocks in Brazil have moderate potential for associated porphyry molybdenum-type deposits. Molybdenite is relatively common in contact zones between felsic volcanic rocks of the Caicara Formation and biotite granite of the Cuchivero Group in Venezuela (Mendoza and others, 1977). In northern Brazil, molybdenite is disseminated in biotite granite and in small quartz veins at the faulted contact between granite and volcanic rocks of the Surumu Formation (Montalvao and others, 1975; Berrange, 1977; Schobbenhaus and others, 1984).

Cassiterite and wood tin are sparsely disseminated in some samples of high-silica rhyolite in the Caicara Formation and its equivalent rocks of the Iricoumé Formation in Brazil (Jones and others, 1986; Sidder, this volume). Cassiterite is also present in panned concentrates such as those from creeks that drain into the upper Río Paragua near the Brazil-Venezuela border (Sidder, this volume, pl. 1). These occurrences of cassiterite disseminated in rhyolite and in panned concentrates are typical of rhyolite-hosted tin deposits (Duffield and others, 1990); however, occurrences of economic significance are not yet known. In Brazil, granitic rocks, such as alaskite, granite, granodiorite, and quartz diorite, of the Uatumã Supergroup that are associated with volcanic rocks equivalent to the Caicara Formation host stockwork veins and disseminations of cassiterite (Damasceno, 1988). The granitic rocks are commonly greisenized to an assemblage of muscovite, fluorite, topaz, and tourmaline (Damasceno, 1988).

The carbonatite at Cerro Impacto (within the Cuchivero terrane) (Sidder, this volume, pl. 1) near the intersection of large northeast- and northwest-striking fractures is enriched in niobium, thorium, barium, cerium, and other metals and rare earth elements (Aarden and others, 1978; Premoli and Kroonenberg, 1981). The northwest-striking fractures may be coextensive with those along which kimberlite was emplaced in the Quebrada Grande area, and they are parallel with large regional fractures that apparently controlled emplacement of pegmatitic dikes into the Parguaza Granite. These fractures extend throughout the western part of the Guayana Shield in Estado Bolívar, Estado Amazonas, and into Brazil. Mendoza and others (1977) suggested that the carbonatitic complex intruded plutonic rocks of the Cuchivero Group during the Mesozoic between 150 and 80 Ma; however, a rubidium-strontium whole-rock isochron date of about 1,732±82 Ma for leached kimberlitic rocks in the Quebrada Grande area (Nixon and others, 1992) indicates that the carbonatite at Cerro Impacto may be much older than Mesozoic. The carbonatite at Cerro Impacto may be Early Proterozoic, or about the same age as kimberlite in the Quebrada Grande area. The date for kimberlite in the Quebrada Grande area is similar to carbonatite emplaced elsewhere in the world between about 1,800 and 1,650 Ma (Meyer, 1988). Additional work is required to date more precisely the age of intrusion of both carbonatite and kimberlite in the Guayana Shield.

The similarity in age, composition, and tectonic environment between the Early Proterozoic Cuchivero Group in Venezuela and the granite-rhyolite terranes of the St. Francois Mountains and the Olympic Dam area suggests that Olympic Dam-type deposits are a possible exploration target in the Venezuelan Guayana Shield (Sidder, this volume). Olympic Dam-type iron-copper-uranium-gold-rare earth element magmatic-hydrothermal deposits are genetically related to magmas, especially those of intermediate to mafic composition, that formed the host granite-rhyolite terrane (Sidder and Day, 1993; Sidder and others, 1993).

UNDIVIDED PROTEROZOIC ROCKS

Many Proterozoic and possibly Archean rocks in Estado Amazonas either have not been studied or have been examined only in reconnaissance fashion and thus are herein included as an undivided group of rocks (fig. 2). These rocks are predominantly granitic and associated volcanic rocks, mafic and alkaline intrusive rocks, and medium- to high-grade gneiss of both igneous and sedimentary protoliths (Mendoza and others, 1977; Gaudette and Olszewski, 1985; Cox, Wynn, and others, 1993). Cordani and Brito Neves (1982) and Goodwin (1991) referred to the granite and high-grade gneiss terrane as basement rocks of the Pakaraima nucleus and considered this nucleus to be an Archean crustal remnant. Mendoza and others (1977) and Gaudette and Olszewski (1985) informally named some of these rocks after the area where they were mapped, such as the Minicia gneiss or migmatite, the Macabana augen gneiss, the granites of Atabapo, San Carlos, Sipapo, and so on. Barrios and others (1985) grouped these locally named units into two provinces, the Ventuari and Casiquiare dominions (fig. 2), on the basis of similar structural, petrologic, and geochronologic characteristics. The Ventuari dominion, primarily north and east of the Río Orinoco in Estado Amazonas, consists of volcanic and plutonic rocks similar to the Cuchivero Group, the Parguaza Granite, sedimentary rocks of the Roraima Group, isolated metasedimentary sequences, and massive alkaline and mafic intrusions. Topographic relief in the Ventuari dominion is high, and elevations are as much as 2,000 m above sea level on the tops of some vertically cliffed plateaus or table mountains, which are called tepuis. The Casiquiare dominion, generally south of the Río Orinoco in Estado Amazonas (fig. 2), includes granite, gneiss, migmatite, and scarce outcrops of the Roraima Group; volcanic rocks and alkaline or mafic intrusive bodies are not present. Elevations rarely are more than 500 m above sea level in this dominion (Barrios and others, 1985). Similar granitic and metamorphic rocks in southeastern Colombia near the Colombia-Venezuela border were named "Complejo migmatítico de Mitú" (Mitú migmatitic complex) (Priem and others, 1982). Notably, all of these authors commented on the complexity of the relations between the rocks, the poor exposures in the jungle, and the intense weathering of the rocks. Most of the mapped and sampled outcrops form discontinuous and isolated exposures along rivers.

The undivided Proterozoic and possibly Archean intrusive rocks in Estado Amazonas range in composition from biotite granite to tonalite and diorite. They are generally medium to coarse grained, equigranular to porphyritic, and weakly foliated. The plutonic rocks have been moderately deformed by small-scale faults and shears, and cataclastic textures are common. Aplite dikes of more than one intrusive episode cut the plutonic rocks (Mendoza and others, 1977; Priem and others, 1982; Gaudette and Olszewski, 1985). Granitic rocks of the Ventuari dominion, such as the granite of Padamo and unnamed plutonic rocks along the Ríos Orinoco, Ventuari, and Paru, are similar to the granite of Santa Rosalía. Associated volcanic rocks of the El Viejo Formation, the volcanic suite of the Parucito Valley, and similar locally named units such as the felsic to intermediate volcanic rocks of Yaví, Asita, Autana, and others in the Ventuari dominion are probably equivalent to the Caicara Formation or to volcanic rocks associated with the Parguaza Granite (Talukdar and Colvée, 1975, 1977; Mendoza and others, 1977; Gaudette and Olszewski, 1985). Some granitic rocks in the Casiquiare dominion are similar to the Parguaza Granite (Barrios and others, 1985; Gaudette and Olszewski, 1985).

The undivided Proterozoic metamorphic rocks range from poorly foliated and mildly tectonized gneiss to well-foliated gneiss and migmatite having cataclastic texture. Phyllite and quartzite are weakly metamorphosed sedimentary rocks. Granitic gneiss, or metamorphosed plutonic rocks, ranges in composition from granite to granodiorite, tonalite, and diorite. The intensity of metamorphism is as high as the greenschist and amphibolite facies with assemblages of chlorite±muscovite±epidote±chloritoid and plagioclase-hornblende±garnet, respectively (Mendoza and others, 1977; Gaudette and Olszewski, 1985).

Geochronological data indicate that peak metamorphism and magmatism of the undivided Proterozoic and possibly Archean rocks in Estado Amazonas occurred between about 1,860 and 1,730 Ma. Gaudette and Olszewski (1985) correlated this metamorphic and magmatic activity with the Trans-Amazonian orogeny, but it is probably part of a younger unnamed event, as discussed previously. Low initial $^{87}Sr/^{86}Sr$ ratios of about 0.703 for paragneiss imply that the protolith of the metasedimentary rocks was not considerably older. Strongly deformed and metamorphosed plutonic rocks of about 1,860 Ma, in addition to moderately to weakly deformed plutonic rocks of about 1,730 Ma,[5] indicate that plutonism and metamorphism may have been synchronous from about 1,860 to 1,730 Ma. Unmetamorphosed, undeformed granitic rocks of about 1,600 Ma represent postorogenic plutonic activity in Estado Amazonas, and the Parguaza rapakivi granite (fig. 2) typifies magmatic activity at about 1,545 Ma. In general, dates for granitic rocks in the Ventuari dominion are older than for those for rocks to the southwest in the Casiquiare dominion (Barrios and others, 1985; Gaudette and Olszewski, 1985; Gibbs and Wirth, 1986).

Geochronological data from the Amazon Territory of Brazil and the Amazonas region of southeastern Colombia demonstrate ages of metamorphic and magmatic activity similar to those for the Casiquiare dominion of Estado Amazonas of Venezuela. Tassinari (1984) utilized whole-rock rubidium-strontium and lead-lead dating methods to date granitic to granodioritic gneiss and migmatite of the Río Negro–Juruena province that truncates the Maroni-Itacaiunas belt (and the Ventuari dominion) in Estado Amazonas and southeastern Colombia (Cordani and Brito Neves, 1982; Teixeira and others, 1989). He determined that a magmatic arc and new continental crust formed in this province between about 1,750 and 1,600 Ma from magma generated in the upper mantle (initial $^{87}Sr/^{86}Sr$ ratio 0.7030).

Priem and others (1982) used whole-rock rubidium-strontium and uranium-lead zircon data to conclude that the granite gneiss basement of the Amazonas region of southeastern Colombia has a minimum age of about 1,850 Ma. A maximum age of about 1,450 Ma was established for the high-grade metamorphism and resetting of the isotopic systems. The Complejo migmatítico de Mitú formed about 1,560–1,450 Ma by large-scale granitic plutonism during the Parguaza episode and metamorphic reconstitution of rocks having a minimum age of about 1,850 Ma (Priem and others, 1982).

Based on the previously discussed radiometric data, Gaudette and Olszewski (1985) suggested that a tectonic zone or boundary is present along the Río Orinoco south of about lat 4°00' N. This zone marks the contact between the Ventuari and Casiquiare dominions and the Maroni-Itacaiunas and the Río Negro–Juruena mobile belts (Barrios and others, 1985; Teixeira and others, 1989). The evolution of a northeast-facing subduction zone between about 1,900 and 1,450 Ma and a change from compressional horizontal tectonics to tensional vertical tectonics at the end of subduction may account for the geologic relations in the undivided Proterozoic rocks of the Amazon region of Venezuela, Colombia, and Brazil (Priem and others, 1982; Barrios and others, 1985; Gaudette and Olszewski, 1985).

RORAIMA GROUP

The Roraima Group is a generally flat lying (dipping less than 20°) suite of sedimentary rocks deposited in fluvial, deltaic, shallow-marine, and lacustrine or epicontinental environments. It originally covered an area of at least 250,000 km^2, possibly as much as 1,200,000 km^2, that extended about 1,500 km east to west from Suriname to Brazil and Estado Amazonas of Venezuela; rocks equivalent to the Roraima Group are apparently not present in French

[5]This date is recalculated from the original published data. The same decay constant, atomic ratios, and experimental errors were used. The results determined here are 1,730±71 Ma, initial ratio=0.705, and MSWD=73.3; however, the data in the publication (Gaudette and Olszewski, 1985) were reported as 1,793±79 Ma, initial ratio=0.704, and MSWD=18.4.

Guiana and the West African craton (Gansser, 1954; Ghosh, 1985; Cahen and others, 1984; Gibbs and Barron, 1993). The original northern extent of rocks of the Roraima Group is not known. The north-facing frontal scarp of the Roraima Group is apparently erosional, although some structural control is possible. Quartzarenite is the predominant rock type; feldspathic arenite, conglomerate, quartzite, arkose, argillaceous siltstone, shale, jasper, and chert are also present. This suite of rocks is generally 700–1,000 m thick and in some areas is more than 3,000 m thick (Ghosh, 1985; Dohrenwend and others, this volume). The basal contact is not exposed in all places; however, the Roraima Group has been reported to overlie unconformably the Early Proterozoic greenstone-granite terrane, the Caicara Formation, the Parguaza Granite, and undivided Proterozoic rocks in Estado Amazonas (De Loczy, 1973; Ghosh, 1985; Gibbs and Barron, 1993). As noted previously, rocks of the Roraima Group locally lie conformably on volcanic rocks of the Caicara Formation. Rocks that overlie the Roraima Group are not common. Fluvial sedimentary rocks about 30 m thick, possibly Miocene to Pliocene in age, overlie the Roraima Group in parts of Guyana and Brazil (De Loczy, 1973); however, Late Proterozoic or Phanerozoic sedimentary rocks are not known to overlie the Roraima Group in Venezuela (Ghosh, 1985; George, 1989).

Rocks of the Roraima Group in Venezuela form startling, vertically cliffed tepuis that rise 1,000 m or more above the surrounding jungle. The tops of the tepuis are commonly saucer shaped and dip about 20°–45° toward the center of the tepuis (Gansser, 1974; Ghosh, 1985). This topography has been attributed to (1) inverted topography in which the tepuis are on the axes of synclines, (2) isostatic depression, (3) buckling related to vertical basement tectonics, and (4) faults along the hinge zones between the marginal rims and the central plateaus that formed central graben and marginal horst blocks (Gansser, 1974; Ghosh, 1985). Briceño and others (1990) concluded that the tepuis resulted from topographic inversion and are remnants of doubly plunging synclines, the low, eroded areas around the tepuis corresponding to the dismantled anticlines. Karst features such as large caves, shafts, underground streams, karren, and dolines are locally developed on the tops of some tepuis (Szczerban and Urbani, 1974), and several species of plants are native only to the tops of individual tepuis (George, 1989).

Structural features in the Roraima Group generally are broad, open folds and block faults that strike northeast and north-northeast to north-northwest throughout Venezuela and the Guayana Shield (Gansser, 1954; Ghosh, 1985; Briceño and others, 1990). Numerous joint sets cut the Roraima Group; however, joint trends measured in outcrop do not coincide with those measured from aerial photographs or side-looking airborne radar (SLAR) images (Sidder, 1988, unpublished data; Briceño and others, 1990). Dohrenwend and others (this volume) recognized a deformational gradient from south to north in the Gran Sabana area of southeastern Venezuela: relatively tight folds and conspicuous axial planar foliation are present in the lower part of the Roraima Group in the south, and relatively undisturbed, gently dipping to horizontal, unfoliated strata are present in the north. The axial planar foliation imposes a conspicuous ridge and valley topography in the south, whereas in the north less deformed and undeformed rocks underlie the high plateaus (Dohrenwend and others, this volume).

Rocks of the Roraima Group generally are unmetamorphosed; most samples do not show any textural or mineralogical evidence of metamorphism (Ghosh, 1985). Pyrophyllite and andalusite in some rocks are interpreted as metamorphic minerals that formed from very low grade burial metamorphism (Urbani, 1977; Briceño and others, 1990); however, these minerals are also in localized contact metamorphic aureoles around diabasic and gabbroic intrusions of the Avanavero Suite or of the Mesozoic dike swarms (Gansser, 1974; Urbani, 1977; Ghosh, 1985; Gibbs, 1986). Rocks interpreted, on the basis of the presence of pyrophyllite or andalusite, to have been affected by burial metamorphism may in fact have been metamorphosed by an unexposed intrusion. Alternatively, an unknown thickness of rocks, but presumably at least 3 km or more to cause burial metamorphism, may have been completely eroded from the top of the Roraima Group. The matrix of some rocks has been recrystallized to interstitial quartz, fine-grained white mica, and hematite (Gibbs and Barron, 1993).

Reid (1974a) divided the Roraima Group in the Gran Sabana area of Venezuela into four formations, in ascending order, the Uairén, Kukenán, Uaimapué, and Matuaí Formations. Elsewhere in the Guayana Shield, the Roraima Group has been subdivided into lower (generally equivalent to the Uairén Formation), middle (the Kukenán and Uaimapué Formations), and upper (Matuaí Formation) members (Gansser, 1954; Bateson, 1966; Priem and others, 1973; Ghosh, 1985; Gibbs and Barron, 1993). The stratigraphy of the Roraima Group is complex, although it may appear simple in some areas. Units are not continuous—rather they appear to grade both laterally and vertically—and few marker beds have been established for regional correlation in the shield. For example, conglomerate, arkose, and jasperoid tuffaceous rocks, which are distinctive in the Gran Sabana area of Venezuela, Brazil, and Guyana, are not present in tepuis in Estado Amazonas (Ghosh, 1985). Moreover, the stratigraphic sequences in several tepuis of Estado Amazonas do not correlate with each other. The rocks in Estado Amazonas are included in the Roraima Group because of their abundant thick sequences of crossbedded, fine- to medium-grained quartzarenite and feldspathic arenite and lesser interbedded layers of shale. Possibly these rocks are time transgressive with those of the Roraima Group in the Gran Sabana area (Ghosh, 1985).

Alberdi and Contreras (this volume) describe a 128-m-thick sequence of graywacke, siltstone, and shale that underlies the basal sandstone and conglomerate of the

Uairén Formation north of the Gran Sabana area. They suggest that these rocks be included in the Roraima Group as the Urico Formation. In northern Brazil and Guyana, the Urico Formation is apparently correlative with the Wailan Formation, which is composed of silty to sandy shale and argillaceous sandstone and is transgressed by conglomerate of the Roraima Group (Gansser, 1954); however, Gansser (1954) considered the Wailan Formation to be equivalent to the Haimarakka Formation, which is correlative with the Caicara Formation rather than with part of the Roraima Group.

The Uairén Formation consists of about 800–900 m of quartzitic sandstone, conglomerate, and minor shaly siltstone. Thin lenses and beds (from less than 50 cm to about 10 m in thickness) of quartz-pebble and polymict conglomerate and thin beds and laminae of shaly siltstone are intercalated with the sandstone (Reid, 1974a; Reid and Bisque, 1975). Dohrenwend and others (this volume) subdivide the Uairén Formation into a lower member about 600 m thick and an upper member about 100–300 m thick. The former consists of well-sorted, coarse- to medium-grained quartz sandstone that is cross-stratified with trough and festoon crossbeds and is intercalated with conglomerate and shaly siltstone; the latter includes medium-grained sandstone that contains abundant trough cross-stratification and intercalated channel gravels (Reid, 1974a; Reid and Bisque, 1975; Dohrenwend and others, this volume). The lower member is moderately bedded to massive and underlies high cliffs and extensive dip slopes of several cuestas along the southern margin of the Gran Sabana. The upper member forms conspicuously benched scarp slopes and irregular ridges (Dohrenwend and others, this volume).

The Kukenán (also spelled Cuquenán) Formation is composed of sandstone interbedded with siltstone, claystone, and shale. The sandstone is well bedded to massive and fine to medium grained, and the siltstone, claystone, and shale are medium to thin bedded, laminated, and variegated. The formation has a maximum thickness of about 100 m in the Gran Sabana area (Reid, 1974a; Reid and Bisque, 1975; Dohrenwend and others, this volume).

The lower part of the Uaimapué Formation is similar to the Uairén Formation in that it consists of sandstone and conglomerate and interbedded siltstone and mudstone; however, abundant beds of jasper, chert, and arkose characterize the upper part of the formation (Gansser, 1954; Reid, 1974a; Reid and Bisque, 1975; Dohrenwend and others, this volume). The total thickness of the Uaimapué Formation in the Gran Sabana area is about 250 m. Sandstone in the lower part of the formation is fine to coarse grained and pervasively channelled with trough cross-stratification. Clasts in the conglomerate are predominantly quartz pebbles. Red arkose, green and red jasper, and green, red, and gray chert are interbedded in the upper part of the formation (Reid, 1974a; Reid and Bisque, 1975). The arkose contains pyroclastic material, and the jasper contains distinct shards of devitrified volcanic glass (Gansser, 1974; Reid, 1976; Ascanio and others, 1985). The jasper beds, which are about 20 cm thick and are interbedded with siltstone and sandstone forming sequences about 10 m thick, are interpreted as volcaniclastic tuff (Ascanio and others, 1985); they are marker beds for correlation throughout the Guayana Shield and have been radiometrically dated.

The Mataui Formation, the youngest unit of the Roraima Group, forms the vertical cliffs of some tepuis and is dominantly crossbedded, ripple-marked, and massive quartz sandstone and quartzite (Reid, 1974a; Reid and Bisque, 1975). Sandstone in the uppermost part of the formation, as on the tops of the tepuis, is less well cemented, friable, and thinner bedded and contains sandy shale horizons. The total thickness of the unit is 1,000 m or more (Gansser, 1954; Reid, 1974a; Reid and Bisque, 1975).

The Roraima Group in Estado Amazonas, which is best observed in tepuis, includes quartzarenite, feldspathic arenite, and shale and generally is made up of three members; however, the stratigraphic sequence cannot be correlated from one tepui to another (Ghosh, 1977, 1985). The lower member, about 300–500 m thick, consists of thinly graded beds of fine- to coarse-grained, cross-stratified, ripple-marked, laminated quartzarenite, minor quartz wacke, and thin conglomeratic beds (Ghosh, 1977, 1985; Mendoza and others, 1977). The middle member, about 100–200 m thick, includes medium-grained, crossbedded quartzarenite, feldspathic arenite, and argillaceous quartzarenite. Dark-gray to black shale units as thick as 50 m are typical of the middle member. The upper member, which forms prominent cliffs, is composed of 500–700 m of medium- to coarse-grained quartzarenite and lesser feldspathic arenite and lenticular clay beds that have thin carbonate-rich laminations (Ghosh, 1977, 1985; Mendoza and others, 1977). Major units of conglomerate, arkose, or jasper have not been observed in Estado Amazonas (Ghosh, 1977, 1985).

ENVIRONMENT OF DEPOSITION AND AGE

Rocks of the Roraima Group were deposited in fluvial, deltaic, shallow coastal marine, and lacustrine or epicontinental environments such as low-sinuosity river channels and their floodplains, delta distributaries above tranquil interdeltaic lakes, coastal lagoons to interdeltaic bays, nonbarred beaches, and intertidal mud flats (Ghosh, 1985; Briceño and Schubert, 1990). The lateral and vertical distribution of the lithologic units depicts an alluvial-deltaic complex near a wave-dominated, high-energy coastline that had abundant beaches, sandy flats, and subtidal bars and less common mudflats and coastal lagoons (Ghosh, 1985). Cross-stratification, ripple marks, and pebble orientation indicate that the sediments were transported from a source to the northeast, east, and southeast (Gansser, 1954; Keats, 1974; Reid and Bisque, 1975; Ghosh, 1985; Gibbs and Bar-

Table 3. Rubidium-strontium whole-rock isochron dates for volcanic rocks of the Roraima Group, Guayana Shield.
[Number of samples on which age is based is given in parentheses after age. MSWD is mean squares weighted deviation]

Country	Unit	Age (Ma)	$(^{87}Sr/^{86}Sr)_0$	MSWD	Reference
Venezuela	Canaima	1,730±120 (*n*=8)[1]	0.708	11.2	Gaudette and Olszewski (1985).
Venezuela	Santa Elena de Uairén	1,570±83 (*n*=16)[2]	0.721	79.8	Pringle and Teggin (1985).
Suriname	Tafelberg	1,660±27 (*n*=14)	0.708	1.84	Priem and others (1973).

[1]These dates are recalculated from the original published data. The same decay constant, atomic ratios, and experimental errors were used; however, the data in the publication were reported as 1,747±49 Ma, initial ratio=0.708, MSWD=2.84.

[2]These dates are recalculated from the original published data. The same decay constant, atomic ratios, and experimental errors were used; however, the data in the publication were reported as 1,579±18 Ma, initial ratio=0.720, MSWD=19.95.

Table 4. Rubidium-strontium whole-rock isochron dates for diabase of the Avanavero Suite and hornfels formed from its intrusion into the Roraima Group, Guayana Shield.
[Number of samples on which age is based is given in parentheses after age. MSWD is mean squares weighted deviation. N.A. indicates not available]

Country	Unit	Age (Ma)	$(^{87}Sr/^{86}Sr)_0$	MSWD	Reference
Suriname	Dolerite of Avanavero Suite	1,670±18 (*n*=22)	0.704	4.24	Hebeda and others (1973).
Guyana	Dolerite of Roraima intrusive suite	1,640±58 (*n*=8)	0.704	11.9	McDougall and others (1963).
Brazil	Hornfelsed sandstone of the Roraima Group next to a diabase dike	1,990±170 (*n*=3)[1]	0.696	0.078	Basei and Teixeira (1975).
Guyana	Hornfelsed shale of the Roraima Group overlying a diabase sill	1,600±44 (*n*=2)[1]	0.856	N.A.	Snelling and McConnell (1969).

[1]These ages are calculated using Model 1 of York (1969), which assumes that the only cause for scatter from a straight line are the assigned errors; however, the errors for $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ were not cited in the original references. Values of 2.0 percent and 0.1 percent, respectively, were used for the calculations.

ron, 1993). Rocks of the Roraima Group may have been deposited in several basins (fault-block basins?) separated by basement highs (Ghosh, 1985).

The Roraima Group may be as old as about 1,900 Ma and as young as about 1,500 Ma or younger (Ghosh, 1985). Correlation of the Roraima Group throughout the shield is problematic because radiometric dating of felsic pyroclastic rocks interbedded with sandstone of the middle part of the Roraima Group (Uaimapué Formation and its equivalents) and diabase dikes and sills that intrude the Roraima Group yields erratic dates. For example, felsic volcanic rocks have whole-rock rubidium-strontium isochron dates of about 1,730, 1,660, and 1,570 Ma (table 3) near Canaima in the northern Gran Sabana area, in Suriname, and near Santa Elena de Uairén in the southern Gran Sabana area, respectively (Gaudette and Olszewski, 1985; Priem and others, 1973; Pringle and Teggin, 1985, respectively). These dates are analytically indistinguishable (1,650 Ma), given the error and MSWD associated with each date (table 3). Diabase dikes that cut rocks of the Roraima Group throughout the central and eastern parts of the Guayana Shield have been dated by $^{40}Ar/^{39}Ar$ and whole-rock and mineral rubidium-strontium and potassium-argon methods. Mineral and whole-rock rubidium-strontium isochrons for doleritic[6] sills that intrude the Roraima Group in Guyana and Suriname yield dates of about 1,640 Ma and 1,670 Ma, respectively (table 4). The potassium-argon method does not give accurate ages for the diabase intrusive rocks (fig. 4) (McDougall, 1968), and $^{40}Ar/^{39}Ar$ analyses of biotite and plagioclase from one sample of diabase gave dates of about 1,800 and 1,470 Ma, respectively (Onstott, Hargraves, and York, 1984). Direct dating of the Roraima Group yields an age of about 1,650 Ma for the middle member (Uaimapué Formation), which rests on as much as 1,200 m of quartzarenite, and indirect dating of diabase dikes and sills indicates an age not younger than 1,670–1,640 Ma for the lower and middle parts of the Roraima Group (McDougall and others, 1963; Hebeda and others, 1973).

Both discordant and concordant contacts have been mapped between rocks of the Roraima Group and the Caicara Formation, which was deposited between about 1,930 and 1,790 Ma (table 2) (Reid and Bisque, 1975; Amaral and Halpern, 1975; Sidder, unpublished data, 1988). In addition, the contact between the Roraima Group and the Parguaza Granite (1,545 Ma) has been identified as an angular unconformity (Mendoza, 1974; Mendoza and others, 1977). In support of this interpretation, Ghosh (1985, p. 48) noted that "contact metamorphic minerals are conspicuous by their absence in the Roraimas overlying the Parguaza Rapakivi granite." In contrast, other Proterozoic units, such as the Cinaruco Formation in Estado Amazonas (equivalent to the Los Caribes Formation), contain andalusite in a contact aureole as much as 0.5 km from the intrusive contact (Ghosh, 1985). These geologic relations bracket the age of the Roraima Group as being as old as about 1,900 Ma in southeastern Venezuela and as young as about 1,500 Ma or younger in Estado Amazonas.

[6]Dolerite is a British term for diabase. The two terms are used synonymously herein; the mafic intrusive rocks in Guyana and Suriname are generally referred to as dolerite (the original usage) and those in Venezuela and Brazil as diabase.

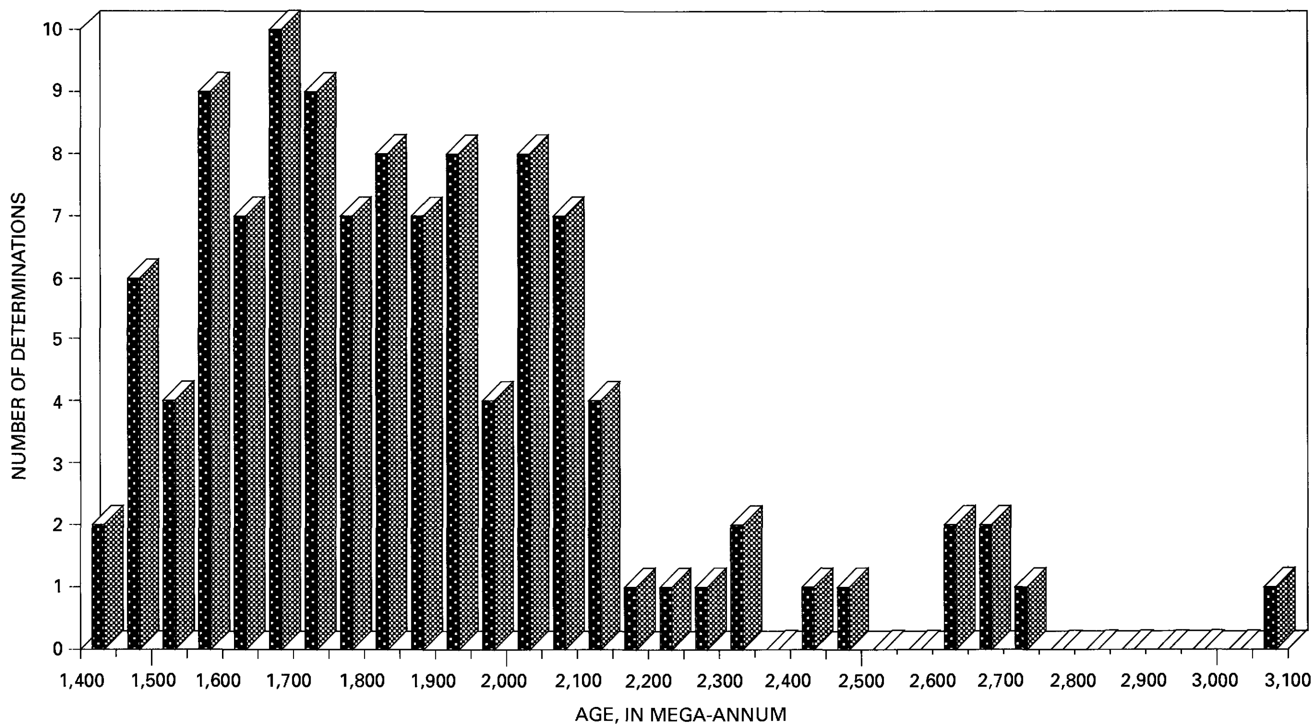

Figure 4. Potassium-argon dates for Proterozoic diabase of the Guayana Shield. Data from McDougall and others (1963), Snelling (1963), Snelling and McConnell (1969), Hebeda and others (1973), Frick and Steiger (1974), Basei and Teixeira (1975), and Teggin and others (1985).

Sedimentary rocks of the Roraima Group apparently record a long period of sedimentation on a relatively stable crust. It has been proposed that the present crustal thickness of shields was not attained at the time of "stabilization," or the cessation of deformation and magmatism (end of the Trans-Amazonian orogeny), but rather that continued passive underplating of the crust (intrusion of part of the Cuchivero Group and its equivalents) and subsequent subsidence and sedimentation may have extended for several hundred million years after initial stabilization (Rogers and others, 1984). The sandstones throughout the Guayana Shield presently correlated with the Roraima Group apparently record a long period of subsidence and continental-nearshore sedimentation that perhaps migrated successively westward in multiple basins on a stable craton possibly from as early as 1,900 to at least 1,545 Ma.

MINERAL RESOURCES

Modern and paleoplacer deposits of gold and diamonds have been mined extensively in the Venezuelan Guayana Shield (Mendoza, 1985; Sidder, this volume). Recent field work (Dohrenwend and others, this volume) in the southern Gran Sabana area identifies three types of placer occurrences: (1) diamond placers within modern channels of major rivers, (2) gold and diamond placers in colluvial-alluvial deposits of low-order drainages, and (3) gold and diamond paleoplacers associated with conglomeratic lenses and beds within the lower 500–600 m of the Uairén Formation. These paleoplacer deposits are the source of gold for modern placer and colluvial-alluvial gravel deposits (Reid and Bisque, 1975; Mendoza, 1985; Dohrenwend and others, this volume). Paleoplacer gold-bearing units in quartz-pebble conglomerate and quartzarenite in the lower part of a fluvatile-deltaic sequence of conglomerate, arenite, quartzite, and intercalated graphitic phyllite in Colombia (Rodriguez and Warden, 1993) are similar to those in the Roraima Group.

Conglomerates in the lower part of the Roraima Group have been proposed as the source of diamonds by many investigators (Gansser, 1954; Pollard and others, 1957; De Loczy, 1973; Briceño, 1984; Mendoza, 1985). Reid (1974b) suggested that kimberlite, possibly in Brazil or even West Africa, was the source for the paleoplacer deposits in the Roraima Group. Studies at San Salvador de Paúl (Briceño, 1984) and elsewhere along the Río Caroní and in the Gran Sabana area document that conglomerates of the Uairén Formation are the source of alluvial diamonds (Reid, 1974b; Reid and Bisque, 1975; Dohrenwend and others, this volume). Briceño (1984) stated that the conglomerates in the Uairén Formation are themselves paleoplacers and were the source for diamond-bearing gravels deposited about 8,000 years ago at San Salvador de Paúl. These Holocene paleoplacers may be the source of diamonds in some deposits now being mined in active drainages (Gansser, 1954; Briceño, 1984).

Kimberlite, a common source of diamonds, and its indicator minerals such as chrome pyrope and magnesian ilmenite have been identified in Venezuela only in the Quebrada Grande area (Baptista and Svisero, 1978; Meyer and McCallum, 1993). More than a dozen diamond-bearing kimberlitic dikes and sills containing chrome pyrope, titanium-rich phlogopite, chromite, and yimengite ($K(Cr,Ti,Fe,Mg)_{12}O_{19}$), a rare alteration product of chromite previously recognized only in kimberlite from China, have been located (Nixon, 1988; Nixon and Condliffe, 1989; Nixon and others, 1989, 1992). A rubidium-strontium whole-rock isochron date of leached kimberlitic samples indicates that the dikes and sills were emplaced at about 1,732±82 Ma (Nixon and others, 1992). Mendoza and others (1977) suggested that the Cerro Impacto carbonatite and possibly associated kimberlite were intruded during the Mesozoic between about 150 and 80 Ma. Although neither kimberlite nor its indicator minerals have been identified elsewhere in Venezuela, it is possible that other Proterozoic diamond-bearing kimberlites were emplaced along the north-northwest-striking fractures that extend throughout the western part of the Guayana Shield in Estado Bolívar, Estado Amazonas, and into Brazil. Sediments of the Roraima Group may have incorporated diamonds from these older Proterozoic kimberlites during their transport and deposition. Numerous diamond-bearing kimberlite intrusions were emplaced in the West African craton during the Mesozoic, particularly about 100 Ma (Williams and Williams, 1977; Morel, 1979). Proterozoic kimberlite intrusions in West Africa are less well documented; however, the Kanangono kimberlite in the Leo Shield of the Ivory Coast is dated as about 1,480 Ma (Onstott and Dorbor, 1987). Alluvial diamonds mined in Ghana and elsewhere in West Africa are presumed to have a source in rocks of the Early Proterozoic Birimian greenstone belts such as tuffaceous graywacke or associated with ultramafic rocks, such as within the Birim diamond field in Ghana (Kesse, 1985; Milési and others, 1992; McKitrick and others, 1993). Neither kimberlite nor kimberlitic indicator minerals has been identified in these rocks.

Radiometric anomalies have been measured in rocks of the Roraima Group on the south end of the Gran Sabana between Santa Elena de Uairén and Icabarú. The apparent similarity of the basal conglomerate of the Roraima Group and gold-uranium-bearing conglomerates of the Witwatersrand, South Africa, Jacobina, Brazil, and Blind River, Canada (De Loczy, 1973; Bellizzia and others, 1981) and the reported occurrence of authigenic pyrite (Gallagher, 1976) have led to the proposal that the basal conglomerates of the Roraima Group are a viable exploration target for uranium deposits, as well as for gold deposits (De Loczy, 1973; Mendoza, 1985; Brooks and Nuñez, 1991). Rocks of the Roraima Group are significantly younger (about 1,900–1,500 Ma), however, than the Early Proterozoic and Archean (3,100–2,200 Ma) gold-uranium deposits and were deposited under a more oxygenated atmosphere. The uranium potential in the Roraima Group is low.

AVANAVERO SUITE

Unmetamorphosed mafic intrusive rocks of the Avanavero Suite (formerly known as the Roraima Intrusive Suite) are present throughout the Guayana Shield as dikes, sills, inclined sheets, and small irregular intrusive bodies such as laccoliths. These rocks are present from western Venezuela to Suriname and Brazil; the age of unmetamorphosed mafic dikes in French Guiana that are cut by Mesozoic diabase dikes is not known (Gibbs, 1986; Gibbs and Barron, 1993). Intrusion of a dense network of mafic dikes at about 1,600 Ma in the Reguibat Shield also indicates a period of contemporaneous crustal extension in the West African craton (Cahen and others, 1984; Rocci and others, 1991). Dikes of the Guayana Shield generally strike north, within 15°, or almost east; they are as thick as 1,000 m and have strike lengths of as much as 150 km (Bosma and others, 1983; Gibbs, 1986). Sills in the Guayana Shield are restricted to the Roraima and Cuchivero Groups, and they were fed by dikes and irregular intrusive bodies in the underlying basement rocks (Hawkes, 1966a; Gibbs, 1986). The sills are commonly intruded along the unconformity between the greenstone-granite terrane and the Roraima Group, as well as throughout the lower and middle members of the Roraima Group (Hawkes, 1966b). The stratigraphically highest sill was previously thought to be at the base of the upper member (Matauí Formation) (Bateson, 1966). On Mount Roraima, at the junction between Venezuela, Guyana, and Brazil, a large gabbroic body forms the boundary between the middle and upper members (Gansser, 1954). Briceño and others (1990) reported, however, that tholeiitic diabase stocks and sills are intrusive into the Matauí Formation in several tepuis of the Chimantá massif in southeastern Venezuela.

Diabasic rocks of the Guayana Shield range from gabbro and norite to granophyre. They are typically medium to coarse grained and massive; chilled margins are developed locally (Hawkes, 1966a; Sial and others, 1986). Subophitic to ophitic textures are common, and the finer grained rocks are commonly porphyritic. Cumulate textures and rhythmic layering are present in the Tumatumari-Kopinang dike and sill complex in Guyana (Hawkes, 1966b). Plagioclase (An_{35}–An_{70}), hypersthene, bronzite, augite, and inverted pigeonite are the primary minerals; accessory amounts of biotite, magnetite, ilmenite, corona-textured olivine, hornblende, and graphic intergrowths of quartz and potassium feldspar (micropegmatite) and trace amounts of apatite, zircon, pyrite, and chalcopyrite are also present (Hawkes, 1966a, b; Hebeda and others, 1973; Gibbs, 1986; Sial and others, 1986; Briceño and others, 1990). The accessory minerals range in abundance from about 2 to as much as 15 modal percent. Minor amounts of secondary minerals such as hornblende and uralite replacing pyroxene, serpentine pseudomorphs of olivine, chlorite after biotite and hornblende, and sericitic and saussuritic alteration of plagioclase are also present. Evidence of metamorphism has not been observed (Hawkes, 1966a, b; Hebeda and others, 1973; Sial and others, 1986). Chemical weathering products of diabase throughout the shield are similar, even at different elevations and on different planation surfaces. Diabase weathers directly to an assemblage of goethite and gibbsite, without intermediate clayey phases (Briceño and others, 1990).

The chemistry of diabase throughout the Guayana Shield is similar to that of continental tholeiite. All samples of dikes and irregular bodies in the basement rocks and sills in the Roraima Group are tholeiitic, and major element concentrations are similar, although trace element abundances of rubidium, barium, yttrium, strontium, and chromium may vary (Hawkes, 1966a, b; Teggin and others, 1985). Gabbroic rocks contain about 51–54 weight percent silica and granophyric rocks about 58–65 weight percent silica (Hawkes, 1966a; Teggin and others, 1985). Diabase analyses plot in the tholeiitic field on several discriminant diagrams such as total alkali-silica, AFM (Na_2O+K_2O-FeO^*-MgO), and Jensen (cation percent (FeO+Fe_2O_3+TiO_2)-Al_2O_3-MgO) plots (Teggin and others, 1985; Sial and others, 1986). Quartz and hypersthene are common normative minerals (Hawkes, 1966b; Teggin and others, 1985). Samples of Proterozoic diabase contain less total iron, titanium, zirconium, vanadium, volatile content (water, carbon dioxide, fluorine, and chlorine as measured by loss on ignition), and possibly copper than samples of the Mesozoic diabase suite (Teggin and others, 1985; Sial and others, 1986). It has been suggested, on the basis of these and other chemical differences between the Proterozoic and Mesozoic diabases, that the Proterozoic rocks represent differentiates of parent magmas that underwent pre-intrusive, shallow-level (crustal) differentiation (Choudhuri, 1978) or that were derived from a shallower region in the mantle (Sial and others, 1986). In contrast, the Mesozoic dike swarms represent magmas derived directly from the mantle, possibly related to a hot spot or mantle plume, during the breakup of Gondwanaland and the formation of the Atlantic Ocean (Choudhuri, 1978; Sial and others, 1986).

Dating rocks of the Avanavero Suite is difficult. As noted previously, diabase from throughout the Guayana Shield has been dated by argon-argon and whole-rock and mineral rubidium-strontium and potassium-argon methods. Potassium-argon dates range from about 3,095 to 1,418 Ma (fig. 4), whereas two mineral and whole-rock rubidium-strontium isochrons yield dates of about 1,670 and 1,640 Ma (table 4). The $^{40}Ar/^{39}Ar$ integrated age dates for biotite and plagioclase in a diabase sill in Guyana are 1,798 and 1,468 Ma, respectively (Onstott, Hargraves, and York, 1984). The last 10 percent of $^{39}Ar_K$ released from plagioclase indicates an age of 1,823 Ma; the last 40 percent of $^{39}Ar_K$ released from biotite defines a plateau at 1,810 Ma, and individual fractions indicate an age of about 1,850 Ma. The dates for biotite are, however, suspect because of the high percentage of atmospheric argon (>29 percent and as much as 91 percent in the fraction at 1,150°C) in 16 of 17 fractions of the biotite analysis. The $^{40}Ar/^{39}Ar$ data for plagioclase support the conclusion that younger dates for the diabase are caused by argon loss (McDougall, 1968; Onstott, Hargraves, and York, 1984).

The lack of a well-defined peak on the histogram of potassium-argon dates for 114 analyses of diabase (fig. 4) indicates that this method has not adequately distinguished the age of the diabasic intrusive rocks of the Guayana Shield. Detailed studies of the distribution of radiogenic argon in samples of diabase prove that the argon is inhomogeneously distributed throughout the rocks and minerals both by excess argon contamination (Hebeda and others, 1973) and by loss of argon (McDougall, 1968) after crystallization of the diabase. These changes are attributed to a tectonothermal event known as the Nickerie or K'Mudku episode at about 1,200 Ma that affected the entire Guayana Shield (Snelling and McConnell, 1969; Hebeda and others, 1973; Bosma and others, 1983). As stated by McDougall (1968, p. 144), "The spread in ages determined by the K-Ar method on the Roraima dolerites shows that little is to be gained by undertaking further measurements by this technique on these rocks. The question as to the age or ages of emplacement of the dolerites probably will only be resolved by detailed and precise Rb-Sr whole-rock measurements on many samples."

Rubidium-strontium isochrons for samples of hornfelsed sandstone and shale collected adjacent to diabase intrusive rocks yield disparate results (table 4). The initial $^{87}Sr/^{86}Sr$ ratio determined for these samples (table 4) indicates that these isochrons may not be representative of the true age of emplacement. The small number of analyses precludes a rigorous interpretation of these data.

Paleomagnetic analyses of Proterozoic diabase from Venezuela, Guyana, and Suriname define at least two groups having remanence orientations approximately opposite in declination (Hargraves, 1968; Veldkamp and others, 1971; Onstott, Hargraves, and York, 1984); however, the radiometric age determinations do not distinguish separate periods of diabase intrusion (table 4, fig. 4) (Gibbs, 1986). At present, the age of diabase intrusion in the Guayana Shield can only be estimated to be about 1,650 Ma, and it may be as old as about 1,850 Ma.

PARGUAZA GRANITE

The Parguaza Granite constitutes a batholith of at least 10,000 km^2, and possibly as much as 30,000 km^2, in Estado Amazonas (fig. 2). These granitic rocks were emplaced predominantly in the Ventuari dominion into rocks equivalent to those of the Cuchivero Group. Granite of similar composition and age is present to the south and southeast into Brazil and westward into Colombia (Mendoza, 1975; Kovach and others, 1976; Gaudette and others, 1978; Priem and others, 1982); these intrusive bodies include the Agua Boa and Madeira plutons, which host the world-class tin deposit at Pitinga, as well as the Surucucus, Mucajai, Abonari, Velho Guilherme, and other intrusive suites in Brazil (Dall'Agnol and others, 1975, 1987, 1994; Dall'Agnol, 1982; Schobbenhaus and others, 1984; Jones and others, 1986; Issler and Lima, 1987; Gibbs and Barron, 1993). Volcanic rocks such as the rhyodacite of Guayapo are associated with the Parguaza Granite (Mendoza and others, 1977). Some felsic to intermediate tuffs in Estado Amazonas identified as correlative with the Caicara Formation may also be related to the Parguaza Granite.

Rocks of the Parguaza Granite include massive, coarsely crystalline, porphyritic granite and biotite granite, commonly having rapakivi (wiborgite-type) texture. They contain quartz (5–34 modal percent), potassium feldspar (microcline perthite, 25–55 percent), plagioclase (oligoclase, 15–31 percent), biotite (3–17 percent), and hornblende (1–24 percent) and accessory clinopyroxene, apatite, sphene, zircon, ilmenite, and magnetite (Mendoza, 1974, 1975; Gaudette and others, 1978). Rapakivi and, less commonly, antirapakivi textures, with ovoids of potassium feldspar mantled by plagioclase and plagioclase mantled by microcline, respectively, are characteristic of these granitic rocks. The Parguaza Granite is relatively unmetamorphosed and unaltered. Epidote is rarely present as a secondary mineral in plagioclase, and chlorite is not present, in contrast to the granitic rocks of Santa Rosalía (Mendoza, 1974; Gaudette and others, 1978).

Rocks of the Parguaza Granite are metaluminous to slightly peraluminous and have a tholeiitic affinity, as indicated by high FeO/(FeO+MgO) ratios. Silica ranges from about 66 to 74 weight percent; amounts of Na_2O and MgO are low to moderate, 2.9–3.9 weight percent and 0.2–0.7 weight percent, respectively; and concentrations of total iron as Fe_2O_3, TiO_2, and CaO are relatively high, 2.6–7.2, 0.4–0.9, and 1.0–3.1 weight percent, respectively (Mendoza, 1975). Major- and trace-element contents of the Parguaza Granite are similar to those of other A-type granites such as granophyre from the Duluth and Skaergaard Complexes, Nigerian charnockite, rapakivi granite in Finland, and Middle Proterozoic (1.48–1.35 Ga) granite in the granite-rhyolite terranes of the Midcontinent of the United States (Mendoza, 1975; Gaudette and others, 1978; Sims and others, 1987).

AGE AND ORIGIN

The Parguaza Granite was intruded about 1.55 Ga in the Ventuari dominion of Estado Amazonas (Gaudette and others, 1978). Other granitic rocks in the westernmost part of the Guayana Shield and southward beneath alluvial cover of the upper Amazon Basin are similiar in age and composition (table 5). Together, these granites have been interpreted to represent widespread anorogenic magmatism at about 1.55 Ga (Dall'Agnol and others, 1975, 1994; Kovach and others, 1976; Gaudette and others, 1978; Priem and others, 1982; Gibbs and Barron, 1983; Teixeira and others, 1989). Anorogenic granite magmatism to the east and southeast in the Central Amazonian Province of Brazil is generally older, having a uranium-lead zircon age of about 1.88 Ga and rubidium-strontium isochron ages of 1.8–1.6 Ga (Dall'Agnol and others, 1994). These older ages suggest that anorogenic granite magmatism in the eastern area was a separate, independent event from that in the west.

The generation of the Parguaza granitic magmas has been characterized as a rift-related event (Gaudette and others, 1978; Gaudette and Olszewski, 1985; Jones and others, 1986; Dall'Agnol and others, 1994). A tectonic environment represented by within-plate crustal extension accompanied by a high thermal gradient due to intrusion of mantle-derived basaltic magmas may explain the origin of the Parguaza Granite. The Parguaza Granite is similar in age, composition, and tectonic setting to the Middle Proterozoic (1,480–1,450 Ma) granite-rhyolite terrane of the St. Francois Mountains in southeastern Missouri of the United States (Kisvarsanyi and Kisvarsanyi, 1989). The formation of rocks in the St. Francois terrane has been explained as a failed cratonic rift or an anorogenic extensional tectonic setting at a passive continental margin (Kisvarsanyi, 1975; Windley, 1989) or, possibly, as the result of shallow subduction, delamination of continental lithosphere, or orogenic-accretionary processes related to the early stages of the adjoining Grenville orogen (Patchett and Ruiz, 1989; Van Schmus, 1993; Storey and others, 1994). The lack of (1) alluvial sediment fill, (2) minor to major quantities of basalt, and (3) basalt erupted peripherally to the rift indicates that the St. Francois granite-rhyolite terrane did not form in a classic rift environment (Patchett and Ruiz, 1989). A similar argument may be made against a rift setting for the Parguaza Granite. Indeed, the Parguaza Granite may be related to shallow subduction or early orogenic-accretionary processes associated with the 1,200-Ma Garzón–Santa Marta granulite belt in Colombia, which has been correlated with the Grenville orogenic belt (Kroonenberg, 1982; Priem and others, 1989).

Ratios such as Na/K, Ba/Sr, and K/Rb indicate that fractional crystallization was an important process during the formation of the Parguaza rapakivi granite (Mendoza, 1974, 1975). The low initial ratio of $^{87}Sr/^{86}Sr$ (0.701, table 5), neodymium isotopic data ($^{143}Nd/^{144}Nd=0.51160$), and high average content of nickel (about 12 ppm, as high as 710 ppm) all suggest that the granitic magmas may have been derived from lower crustal material of trondhjemitic or charnockitic composition and a component of undifferentiated mantle material (Mendoza, 1974, 1975; Gaudette and others, 1978; Allègre and Ben Othman, 1980). Initial ratios of $^{87}Sr/^{86}Sr$ between 0.704 and 0.716 and uranium-lead inher-

Table 5. Rubidium-strontium whole-rock isochron dates for granitic rocks of the Parguaza province and their equivalents, Guayana Shield.

[Number of samples on which age is based is given in parentheses after age. MSWD is mean squares weighted deviation. N.A. indicates not available]

Country	Unit	Age (Ma)	$(^{87}Sr/^{86}Sr)_o$	MSWD	Reference
Venezuela	Parguaza Granite	1,490±120 (n=4)[1]	0.701	2.09	Gaudette and others (1978).
Venezuela	Granite of San Carlos de Río Negro (Casiquiare dominion)	1,567±25 (n=4)	0.704	0.93	Gaudette and Olszewski (1985).
Colombia	Granites of the Río Inírida and Río Guaviare (Ventuari? dominion)	1,485±35 (n=8)	0.706	1.6	Priem and others (1982.
Brazil	Surucucus granite	1,520±140 (n=6)[1]	0.696	22.2	Basei and Teixeira (1975), Dall'Agnol and others (1975).
Brazil	Granite of the Upper Amazon Basin	1,530±25 (n=3)[2]	0.706	0.29	Kovach and others (1976).
Brazil	Agua Boa-Madeira plutons, Pitinga area	1,700±34 (n=9)[3]	0.701	5.98	Macambira and others (1987).

[1]The errors for $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ were not cited in the original references. Values of 1.5 and 0.085 percent (H.E. Gaudette, University of New Hampshire, oral commun., 1990) and 2.0 and 0.1 percent, respectively, were used for the calculations of the dates of the Parguaza and the Surucucus granites, respectively. The Parguaza Granite has a uranium-lead zircon age of 1,545±20 Ma (Gaudette and others, 1978). Dall'Agnol and others (1993) reported a reference isochron age of 1,583 Ma at an initial ratio of 0.708 for the Surucucus granite.

[2]These three samples are several hundred kilometers from one another and may not be comagmatic. Their model ages, assuming an initial ratio of 0.706, are 1,541, 1,536, and 1,528 Ma.

[3]This date is recalculated from the original published data. The same decay constant and atomic ratios were used; however, the date in the publication was reported as 1,689±19 Ma, initial ratio=0.707, and MSWD=1.48, which is apparently a model 1, not a model 3, date.

itance patterns in zircon in similar anorogenic granite from other parts of the shield indicate a crustal source or an important contribution of continental crust to the magmas that generated these rocks (Pimentel and others, 1991; Dall'Agnol and others, 1994).

MINERAL DEPOSITS

Placer, eluvial, and lode occurrences of tin in Estado Amazonas and Estado Bolívar are spatially associated with the Parguaza Granite. Cassiterite in lodes is associated with quartz veins that cut the granite, and anomalous values of tantalum, niobium, zirconium, and titanium, contained in tantalum-rich rutile or struverite, tantalum-niobium-iron-manganese-bearing rutile, tantalite-columbite, stanniferous tantalite or ixiolite, and zircon (Aarden and Davidson, 1977), are present in pegmatite associated with the granite (Rodriguez and Perez, 1982; Perez and others, 1985). The best known tin prospect is that near Caño Aguamena (Sidder, this volume, pl. 1).

The Parguaza Granite is equivalent in composition and origin and approximately equal in age (table 5) to granite in the Agua Boa and Madeira plutons, which host the Pitinga deposit in Brazil. Pitinga is one of the world's largest tin deposits and produces about 12 percent of the western world's tin (Thorman and Drew, 1988). As of 1984, it had measured reserves of about 203,000 metric tons of tin, of which about 30 percent was in alluvial deposits (Daoud and Antonietto, 1988). Measured, indicated, and inferred reserves totaled about 269,000 metric tons of tin; alluvial reserves accounted for about 64 percent (Daoud and Antonietto, 1988). Greisenized granite, locally called apogranite, hosts the primary tin ore and is the source of the alluvial deposits (Jones and others, 1986; Macambira and others, 1987; Daoud and Antonietto, 1988; Thorman and Drew, 1988). In the Surucucus area of northernmost Brazil, alkaline granite having rapakivi texture is about 1.5–1.6 Ga and has potential reserves of 20,000 metric tons of tin (Dall'Agnol and others, 1975, 1994; Schobbenhaus and others, 1984; Jones and others, 1986). Thus, the Parguaza Granite has high potential for identification of undiscovered tin deposits.

The similarity in age, composition, and tectonic environment between the Parguaza Granite and the granite-rhyolite terrane of the St. Francois Mountains suggests that Olympic Dam-type iron-copper-uranium-gold-rare earth element deposits are a favorable exploration target in the Parguaza terrane (Sims, 1988; Sidder and Day, 1993; Sidder and others, 1993). Aeromagnetic and radiometric data could help locate potential prospects. A highly magnetic area and a superimposed uranium anomaly detected in aeromagnetic and aeroradiometric data over the Parguaza Granite in the westernmost part of Estado Bolívar (lat 5°10′ N., long 64°20′ W.) (Wynn, 1993) is a good area to explore further.

The bauxite deposit at the Los Pijiguaos mine formed from the Parguaza rapakivi granite (Moreno and Bertani, 1985b). The richest ore is at an erosional level (known as the Imataca-Nuria erosion surface) at elevations between 620 and 690 m, and it formed during an intense weathering cycle in the Late Cretaceous and early Tertiary (Short and Steenken, 1962; Menendez and Sarmentero, 1985; Schubert and others, 1986). Bauxita Venezolana C.A. (BAUXIVEN) produced about 245,157 metric tons of ore in 1987 during its first year of operation at Los Pijiguaos, almost 2 million metric tons in 1991, and a reported 1.05 million metric tons in 1992 (Ensminger, 1992; Doan, 1994). Initial measured and indicated reserves of bauxite were 201.8 million metric tons with a grade of 48.7 percent Al_2O_3 and 10.9 percent SiO_2, which included reserves of 70.1 million metric tons with 51.8 percent Al_2O_3 and 6.4 percent SiO_2 (Menendez and Sarmentero, 1985). Newman (1989) reported that proven (200 million metric tons) and probable (500 million metric tons) reserves of 700 million metric tons are present in the Los Pijiguaos area. Three new bauxite deposits in the Parguaza terrane were discovered in 1989 between the Los Pijiguaos mine and Puerto Ayacucho on the Río Orinoco (Ensminger, 1992).

NICKERIE OROGENY

Potassium-argon, argon-argon, and rubidium-strontium dates of about 1,350–1,100 Ma for mica and feldspar from Archean and Early Proterozoic rocks of the Guayana Shield are indicative of partial resetting and overprinting due to the Nickerie metamorphic episode (Priem and others, 1968; Kroonenberg, 1982; Onstott and others, 1989). Loss of argon and strontium due to recrystallization resulted in the abnormally young Middle Proterozoic dates for Early Proterozoic and Archean rocks, and increase of argon in diabase of the Avanavero Suite caused the aberrantly old dates (fig. 4). Rocks affected by this episode extend from western Suriname through Guyana to Venezuela, Colombia, and northern Brazil. The eastern boundary of reset mica ages is in central Suriname (Priem and others, 1971; De Vletter and Kroonenberg, 1984). East of this boundary, Early Proterozoic rocks show Trans-Amazonian, not Nickerie, mineral ages. In Venezuela the Nickerie orogeny is also called the Orinoquean orogenesis (Mendoza, 1977a; Moreno and others, 1977); in Guyana it is named the K'Mudku mylonite episode (Barron, 1969; Singh, 1974); and in Brazil it is known as the Jari-Falsino event (Kroonenberg, 1982). The Nickerie orogeny is equivalent to the Grenville orogeny in North America and Antarctica (Moores, 1991; Storey and others, 1994).

Reactivation of east-northeast-striking faults, such as the Guri shear zone, and minor uplift throughout the central and western parts of the Guayana Shield characterize the Nickerie episode. Cataclastic textures and locally mylonite zones and pseudotachylite developed along some faults (Short and Steenken, 1962; Priem and others, 1968; Barron, 1969; Gibbs and Barron, 1993). Minor aplite and pegmatite dikes may have been emplaced coincident with faulting (Mendoza, 1977a). Extremely low grade to medium-grade metamorphism due to cataclasis and mylonitization in the central Guayana Shield during the Nickerie episode produced minerals such as pumpellyite, prehnite, epidote, albite, muscovite, chlorite, biotite, stilpnomelane, sphene, actinolite, and garnet (De Roever and Bosma, 1975). In the western part of the shield, high-grade metamorphism generated charnockitic and enderbitic granulite, mafic granulite, amphibolite, and augen gneiss (Priem and others, 1989).

The Nickerie metamorphic episode commonly has been described as a regional tectonothermal event (Priem and others, 1968; Barron, 1969; De Roever and Bosma, 1975; Mendoza, 1977a), but a cause for the tectonism or increased heat flow has not been identified. Recent geochronological studies of the Garzón massif in the Andes of Colombia indicate that a quartzofeldspathic, calc-alkaline (continental arc) sequence of rocks along the western margin of the Guayana Shield was metamorphosed to granulite facies about 1,172 Ma (Priem and others, 1989). This metamorphism and associated deformation are attributed to continental collision (Kroonenberg, 1982; Priem and others, 1989; Park, 1992). This collision resulted in an early Neoproterozoic supercontinent during the 1.3–1.0-Ga Grenvillian (-Nickerie) orogeny (Dalziel, 1992; Storey, 1993). The reset mineral ages in rocks of the Guayana Shield are herein interpreted to be the result of thermal and tectonic effects in the hinterland of this proposed collision during the Nickerie orogeny.

MESOZOIC DIABASE DIKES

Narrow (<200 m), thin (<50 m), long (as much as 250 km) unmetamorphosed diabase dikes that trend approximately east-northeast in Venezuela and north-northwest in the eastern part of the Guayana Shield are related to the opening of the Atlantic Ocean (MacDonald and Opdyke, 1974; Gibbs, 1986). They have been called the Apatoe dike suite in Guyana (Gibbs and Barron, 1993). These dikes are generally thinner and straighter than dikes of the Avanavero Suite, possibly due to a more rigid crust at the time of their emplacement (Gibbs and Barron, 1993); however, because the field appearance and chemistry of these diabase suites are similar, the Mesozoic dikes are grouped with those of the Avanavero Suite (unit Xa) in figure 2. The dike rocks are fine to medium grained, have subophitic to ophitic texture (Hargraves, 1978), and contain plagioclase (commonly labradorite) and augite, minor pigeonite, relict olivine cores in pyroxene, biotite, green amphibole, apatite, opaque minerals such as titaniferous magnetite, ilmenite, and rare chalcopyrite and pyrrhotite, and minor interstitial granophyric intergrowths of quartz and microcline microperthite (Hawkes, 1966a; Hargraves, 1978; Choudhuri and others, 1984). These rocks are quartz-saturated tholeiite that has a continental basalt affinity (Choudhuri and others, 1984), and they generally contain more titanium, total iron, zirconium, vanadium, volatile content (water, carbon dioxide, fluorine, and chlorine as measured by loss on ignition), and possibly copper than does diabase of the Avanavero Suite (Choudhuri, 1978; Teggin and others, 1985). The chemistry of the Mesozoic rocks suggests that the dike swarms were derived directly from an undepleted mantle source, possibly related to a mantle plume and hot spot, during the breakup of Gondwanaland and the separation of South America from Africa (Choudhuri and others, 1984).

The age of these younger diabase dikes, which are present throughout the Guayana Shield, has not been determined precisely. As shown in the histogram in figure 5, potassium-argon dates for 59 samples of diabase range from about 550 to 130 Ma. The majority of dates are between about 230 and 170 Ma, and many are about 210 Ma (fig. 5). Dikes that have different potassium-argon dates have similar magnetic poles that are poles characteristic of Permian-

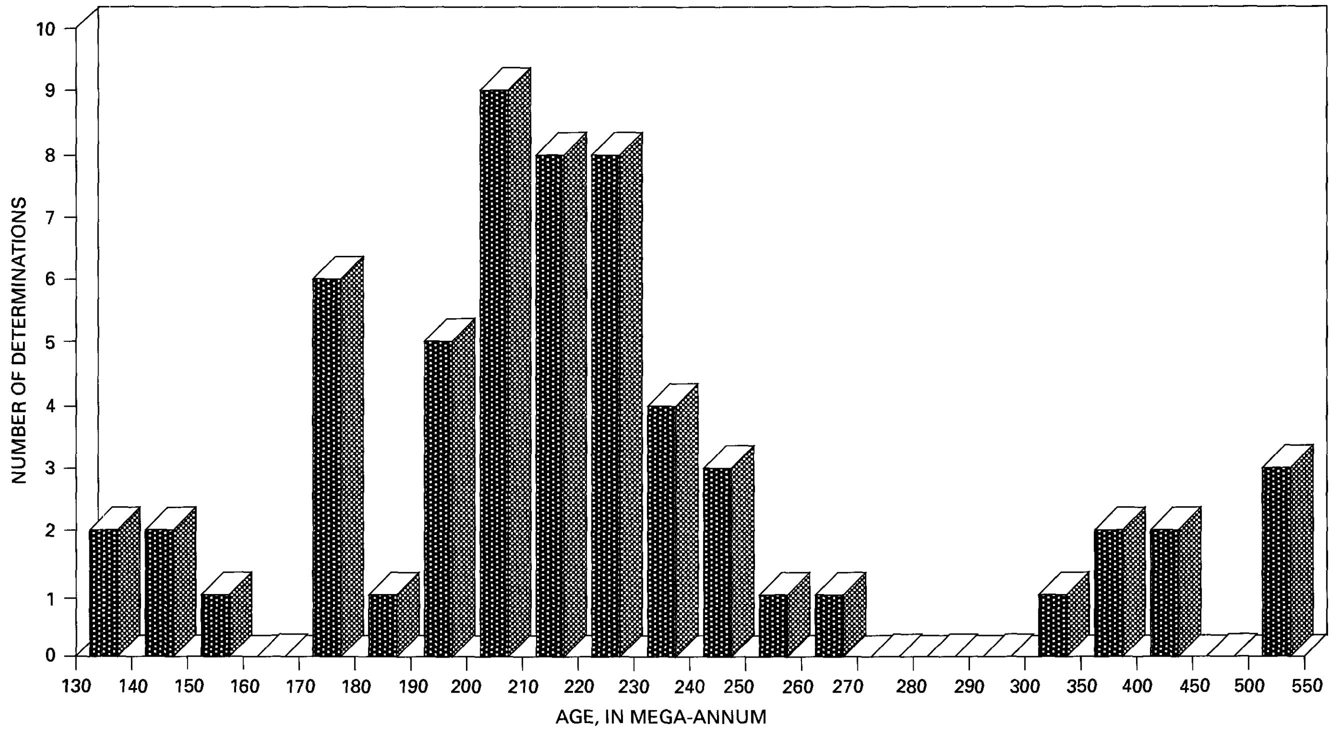

Figure 5. Potassium-argon dates for Phanerozoic diabase of the Guayana Shield. Data from Priem and others (1968), Frick and Steiger (1974), MacDonald and Opdyke (1974), Schobbenhaus and others (1984), and Teggin and others (1985).

Triassic rocks (Veldkamp and others, 1971; Hargraves, 1978); however, scatter of the paleomagnetic pole positions does not allow the age of the dikes to be distinguished more accurately. Dates for diabase dikes in Liberia, West Africa, that intruded Proterozoic basement rocks range from about 1,222 to 177 Ma. These dikes contain large and variable amounts of excess ^{40}Ar that resulted in anomalously old dates (Dalrymple and others, 1975; Mauche and others, 1989). Dates for diabase dikes that intruded Paleozoic sandstone range, however, from about 201 to 177 Ma. Dikes that cut sandstone may also contain small amounts of extraneous ^{40}Ar (Dalrymple and others, 1975; Mauche and others, 1989). Thus, dates for diabase dikes in the Guayana Shield, all of which intrude Proterozoic or Archean rocks, may not be representative of the time of dike emplacement and crystallization. All of the dikes are probably latest Triassic to earliest Jurassic, about 210–200 Ma, in age. It is not known whether emplacement of the dikes marked the initiation of or predated the separation of South America and Africa and the opening of the Atlantic Ocean.

MESOZOIC-CENOZOIC UPLIFT, EROSIONAL SURFACES, AND ALLUVIUM

Uplift and southward tilt of the Imataca Complex and other rocks in the northern Guayana Shield of Venezuela occurred just prior to or during the separation of South America from Africa (Onstott and others, 1989). Additional uplift and tilt may also have been associated with mid-Cenozoic orogeny in the Caribbean area (Short and Steenken, 1962; Olmore and Estanga, 1989; Olmore and García-Gerdes, 1990; Olmore and others, 1993). These uplifts caused erosion of the Guayana Shield and subsequent deposition in a basin north of the Río Orinoco (Schubert and others, 1986; Olmore and García-Gerdes, 1990; Olmore and others, 1993). Geomorphic and geologic evidence of Cenozoic tectonism has not been observed in the Gran Sabana area; however, slow, broad regional upwarping may not have generated a recognizable geomorphic expression (Dohrenwend and others, this volume).

At least six planar geomorphic surfaces, marked by distinct elevations, have been identified in the Venezuelan Guayana Shield. They are, from oldest to youngest, (1) Auyantepui (2,000–2,900 m), (2) Wonken or Kamarata-Pakaraima (900–1,200 m), (3) Imataca-Nuria (600–700 m), (4) Caroní-Aro (200–450 m), (5) Llanos (80–150 m), and (6) Orinoco floodplain (0–50 m) (Short and Steenken, 1962; Menendez and Sarmentero, 1985; Schubert and others, 1986, 1989; Briceño and Schubert, 1990; Gibbs and Barron, 1993; Dohrenwend and others, this volume). The ages of the oldest two surfaces are not well known; Schubert and others (1986) speculated that they are Mesozoic. The other surfaces range in age from early Tertiary to Holocene (Schubert and others, 1986; Briceño and Schubert,

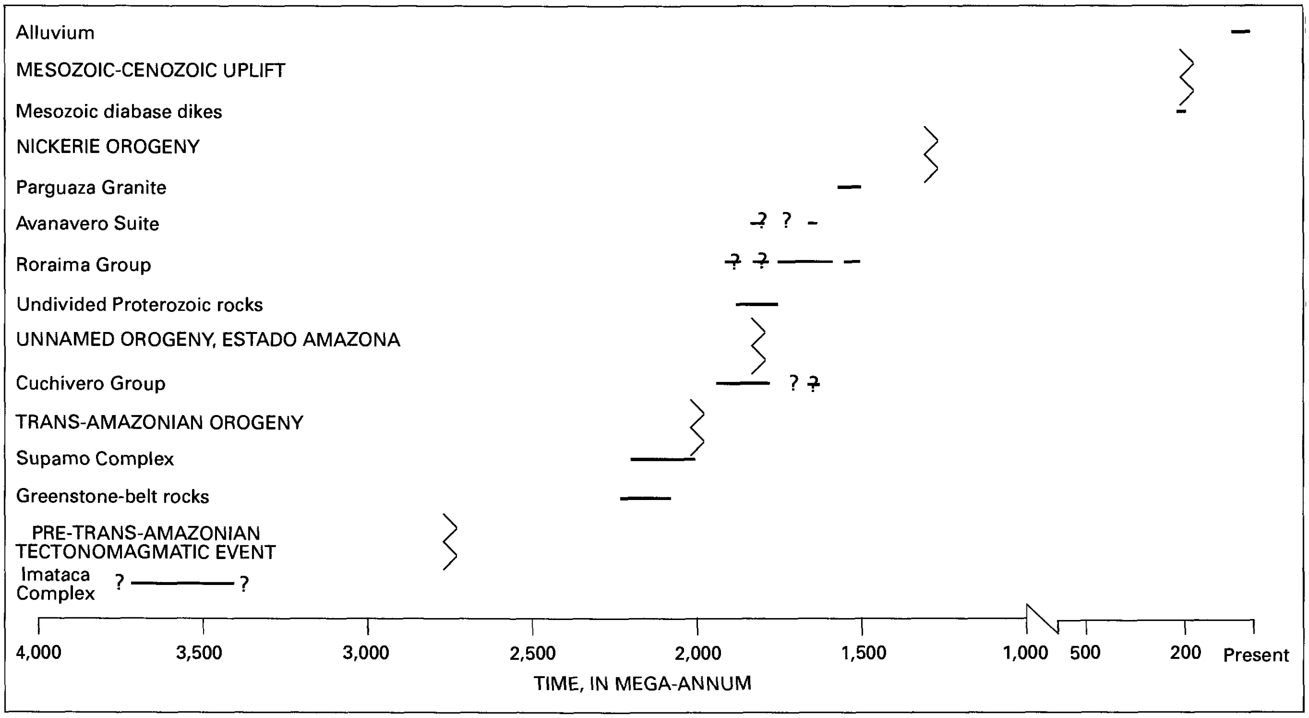

Figure 6. Chronology of the Venezuelan Guayana Shield. Rock units described in table. 1

1990). The Eocene-Oligocene Imataca-Nuria surface is important economically because bauxite and enriched deposits of banded iron formation are developed on it (Menendez and Sarmentero, 1985; Schubert and others, 1986). All of the planation surfaces have been correlated with similar surfaces in Brazil, Guyana, Suriname, and (or) French Guiana, as well as in West Africa and India (Short and Steenken, 1962; Prasad, 1983; Schubert and others, 1986; Briceño and Schubert, 1990; Gibbs and Barron, 1993).

Episodic periods of tectonic uplift, landscape dissection in response to uplift, and parallel slope retreat during succeeding intervals of stability are considered to be the cause of formation of these planation surfaces. Many of the surfaces coincide, however, with lithologic units, and the steps that separate them commonly coincide with lithologic discontinuities (Kroonenberg and Melitz, 1983). Furthermore, within the southern part of the Gran Sabana, ferricrete-capped remnants of a middle(?) Tertiary geomorphic surface cut across several of the regional planation surface levels and extend almost continuously from dissected strath terraces lying less than 10 m above modern streams to high, structurally controlled ridges rising more than 200 m above the floors of the larger valleys (Dohrenwend and others, this volume). Thus, present-day relief within the Guayana Shield is mostly a function of lithologic resistance. At least some of these so-called planation surfaces are a manifestation of lithologic control in a relatively stable landscape that is developed on a thick sequence of flat-lying to very gently dipping sedimentary strata.

Tertiary-Quaternary paleoplacer deposits and the lower Roraima Group are the source of diamonds and gold in Holocene alluvium (Briceño, 1984; Dohrenwend and others, this volume).

SUMMARY

The Guayana Shield of Venezuela consists predominantly of Archean and Early to Middle Proterozoic rocks (fig. 6). Archean rocks are not known elsewhere in the Guayana Shield, but similar Archean rocks and tectonic events are present in the West African Shield. Early Proterozoic greenstone-belt rocks and orogenic events in the rest of the Guayana Shield and in the West African Shield are comparable to those in the Venezuelan Guayana Shield. Differences in the types of late Early Proterozoic and Middle Proterozoic rocks in the two shields are notable. For example, granite and rhyolite equivalent to the Cuchivero Group are present throughout the Guayana Shield but are only locally present in the West African Shield. Also, rocks equivalent to the Roraima Group and the Parguaza Granite, although widespread in the Guayana Shield, are absent or rare in West Africa.

The only confirmed Archean rocks in the Guayana Shield are in the Imataca Complex. The Imataca Complex includes gneiss, granulite, amphibolite, dolomite, manganiferous rocks, and banded iron formation. Amphibolite- to granulite-facies metamorphism and refolded isoclinal folds are characteristic of the Imataca Complex.

The Early Proterozoic greenstone belts in Venezuela comprise submarine sequences of mafic volcanic rocks at the base, a middle section of basalt to rhyolite, and an upper section of turbiditic graywacke, volcaniclastic rocks, and chemical sedimentary rocks. Tholeiitic and calc-alkaline differentiation trends are common in the volcanic rocks. Layered mafic complexes are also present in the greenstone belts. Rocks of the greenstone belts formed between about 2,250 and 2,100 Ma. They were metamorphosed to the greenschist facies and locally the amphibolite facies near granitic domes of the Supamo Complex, which intruded the greenstone belts between about 2,230 and 2,050 Ma. Major deposits of low-sulfide gold-quartz veins are hosted by rocks of the greenstone belts in shear zones. The Imataca Complex and the greenstone-granite terranes were deformed and metamorphosed during the Trans-Amazonian orogeny, which represents a period of continental collision between about 2,150 and 1,960 Ma. Postcollisional, post-Trans-Amazonian magmatism between about 1,930 and 1,790 Ma produced volcanic and plutonic rocks of the Cuchivero Group. Kimberlite in the Quebrada Grande area (and possibly carbonatite in the Cerro Impacto area) was intruded about 1,730 Ma. Undivided Proterozoic and possibly Archean(?) rocks in Estado Amazonas include granitic rocks, gneiss, and migmatite. Peak metamorphism and intrusion of these rocks occurred between about 1,860 and 1,730 Ma.

Unmetamorphosed rocks of the Roraima Group were deposited in fluvial, deltaic, shallow coastal marine, and lacustrine or epicontinental environments. The Roraima Group is possibly as old as about 1,900 Ma and as young as about 1,500 Ma. Conglomeratic lenses and beds in the lower 500–600 m of the Roraima Group contain paleoplacers, some of which are the source of diamonds and gold in modern placer deposits. Mafic dikes, sills, and irregular intrusive bodies of the Avanavero Suite cut all earlier rocks of the Guayana Shield. These continental tholeiitic intrusions are about 1,650 Ma in age and possibly as old as about 1,850 Ma. Middle Proterozoic, 1.55-Ga rapakivi granite of the Parguaza terrane hosts occurrences of tin in quartz veins. Major deposits of bauxite are developed on the Parguaza Granite. Continental collision during the Nickerie orogeny in the westernmost part of the Guayana Shield reset mineral ages in Archean and Early Proterozoic rocks of the central and eastern parts of the shield to about 1,200 Ma. The Nickerie orogeny is equivalent to the Grenville orogeny in North America.

Diabase dikes intruded the Guayana Shield during the opening of the Atlantic Ocean about 210–200 Ma. Kimberlite and carbonatite may have been intruded during the Mesozoic. Six planar geomorphic surfaces developed at distinct elevations in the Guayana Shield during the Mesozoic and Cenozoic eras. Deposits of bauxite and enriched banded iron formation formed on the early Tertiary Imataca-Nuria surface. Gold and diamond placers are present in modern major river channels and in colluvial-alluvial deposits of low-order drainages.

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