OKAYAMA UNIVERSITY

Earth Science Reports

Vol. 16, No. 1, 2009

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Okayama University

Earth Science Reports

Vol. 16, No. 1, 2009 (Annual)*

Publisher
Department of Earth Sciences, Faculty of Science,
Okayama University
1-1, Tsushima-Naka 3chome, Kita-ku, Okayama 700-8530, JAPAN
TEL. 81-86-251-7891 FAX. 81-86-251-7895
http://www.desc.okayama-u.ac.jp/

* No. 1 is issued annually, additional volumes of the same year
will be published occasionally using No. 2, etc.

Printed in Japan

OKAYAMA University
Earth Science Reports,
Vol. 16, No. 1, 1-7, (2009)

日本最初の用語「地質学」の成立: 箕作阮甫(1799-1863)の貢献

The first Japanese word "chishitsu-gaku" for the "geology" was proposed by Genpo Mitsukuri (1799-1863)

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The first Japanese word "chishitsu-gaku" for the "geology" was proposed by Genpo Mitsukuri, a scientific contributor in the Tokugawa regime in the 19th century, who was born in 1799 in Tsuyama in Western Honshu, Japan and had worked for scientific activities of the Tokugawa government from 1839 to 1863. His major works for the first making of the "chishitsu-gaku" for the "geology" have been presented in this paper, also showing his contribution to Japanese sciences.

Keywords: Genpo Mitsukuri, "chishitsu-gaku", 19^{th} century

図 1 箕作阮甫肖像（津山洋学資料館にて．岡田博有撮影）

I はじめに

日本の地質学の一般的な表示は今井 功や土井正民などによって示されたが（今井，1966，1970; 土井，1978; 岡田，2002; Okada with Kenyon-Smith, 2005），その具体的な表現の仕方について筆者は帆足万里（ほあし ばんり）の貢献に敬意を評した（岡田，2008）。しかし，そのとき分かったことは，日本の用語「地質学」はそれまで帆足万里も重要な貢献をしていたと思われていたが，実際には帆足万里は「地質学」用語の提唱はしていなかったのである（岡田，2008）。

そこで，帆足万里に続いて活躍を始めた蘭学者箕作阮甫（みつくり げんぽ）が具体的な「地質学」用語の作成を行い，その重要な成果をはたしていたことをここで明らかにしたい。現在の岡山県津山市にあった津山藩で活躍を始めた箕作阮甫は，当時の日本政府の中心であった江戸の幕府に移って蘭学者としての活動を行

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い，幕府の重要な科学者としての研究を行った。その貴重な成果が，19 世紀の日本の自然科学発展の中で始めて「地質学」の重要性を示したことである。その実態をここで明らかにして,「地質学」用語が箕作阮甫によって作られたことを示したい。なお，後閑（1970）が，日本における近世前の化石の研究は進んでいたが，さらに箕作阮甫が 1861（文久 元）年に「地質辦証」書を訳述して地質に関する地史を述べた，と解説している。こうして作られた日本の地質学は明治時代に最初の専門分野として大発展を遂げ（矢部，1953)，その地質学の研究と教育の機関が 1877（明治 10）年に東京帝国大学理学部に創設されたのである（坪井，1953; 土井，1978)。さらに驚いたことに，我々もよく知っていた東京大学地質学科教授坪井誠太郎（つぼい せいたろう）が箕作阮甫の子孫であったことである。我々が坪井誠太郎の名前をよく知っているのは，偏光顕微鏡による固態物質の光学的研究法を解説した「坪井誠太郎（1959）: 偏光顕微鏡．岩波書店」からであった。さらに，日本の地質学の発展に大貢献をした矢部長克（矢部，1970）はまた坪井誠太郎の父であった坪井正五郎（東京大学教授，人類学）にもふれている。

また，日本の地学教育について，1890（明治 3）年以後の東京における中学校教育では，箕作阮甫の「地球説略」が取り上げられている（倉林，2000)。

こうして大発展を遂げた日本の地質学の形成が，箕作阮甫による新用語「地質学」作成への貢献を基にしていることを，ここに明らかにしたい。

Ⅱ 箕作阮甫の生い立ちと自然科学との関係

箕作阮甫の生い立ちと自然科学への貢献についての資料は，主として菊池（1978)，小沢（1978)，児玉（1978)，緒方（1978)，大久保（1978)，土井（1978)，小山（1978)，有坂・西川（1985)，津山洋学資料館（1990，2008)，木村（1994）などを参考にしている。

箕作阮甫（みつくり げんぽ)（図 1，2; 表 1）は 1799（寛政 11）年 9 月 7 日に津山（つやま）（現在の岡山県津山市）で生まれた。名は虔儒（けんじゅ)，字は庠西（しょうせい)，阮甫は道称であったが（緒方，1978)，1828（文政 11）年には正式名となった（小山，1978)。さらに，箕作阮甫には津山の地名にちなんで紫川，逢谷の号があった。

この津山は 1603（慶長 3）年に 18 万 6500 石で入封した森忠政が新城を築いた後，松平宣富が津山十万石の城主になり，その後五万石の城主のときの 1799（寛政 11）年に箕作阮甫の誕生となった。この時期は，洋学が盛んになったときで,1801年には19世紀の時代が始まった。この時期の日本では文化，文政，天保，弘化，嘉永，安政，文久と年代が変わり，19 世紀初期の日本は政治，社会の動きが極めて重要な時代であった。特に，外国との関係が重要で，これに合わせるように長州藩や薩摩藩の活躍が大きく，大成功を続けていた。その中で，とくに鹿児島の薩摩藩藩主の島津斉彬（しまず なりあきら）は西洋式で箕作阮甫と深い関係があったようだ（児玉，1978)。

箕作阮甫の両親としての父は箕作丈庵（貞固)（1759-1802)，母は清子であった。父は箕作家三代目の医師として評判が高く，津山藩の侍医も勤めていた。 箕作阮甫はその次男として，すでに述べたように，1799（寛政 11）年 9 月 7 日に津山の西新町（にししんまち）で代々町医師を営む家に生まれた。しかし，阮甫が 4 歳の時の 1802（享和 2）年に父「丈庵」が 44 歳で亡くなった。その後，まもなく 1810（文化 7）年に兄の「豊順」も 17 歳で没し，彼の子供もいなかったので，箕作阮甫が 15 歳のとき箕作家を継いだ。さらに,阮甫が 25 歳のときの 1823（文政 6)年 9 月 25 日には母清子が亡くなった。

箕作阮甫は幼いときに怪我（けが）を患い，右の肘関節（ひじかんせつ）が利かなくなった。このため，阮甫の母は彼を武士にしないで，医師の道を選ぶことにさせた。

こうして，箕作阮甫は 1815（文化 12）年の 17 歳のときに,京都で吉益文輔の指導で漢方医学の学業を始めた。この京都での箕作阮甫の修業には藩から年間 5 両が給付されていた。こうして蘭学者箕作阮甫の生まれる素地ができたのであった（小山，1978)。予定通り三年後の

1819（文政 2）年 21 歳のときに阮甫は京都から津山に帰った。そして，この年の 11 月に阮甫は同藩の養女「大村とい」（大村登井）と結婚，阮甫は箕作家の学問的な根となった（小山，1978）。この結婚で 3 人の女子が生まれた。この 3 人の子女［木村（1994）に依ると，実際は 4 人の子女であったが，一人の死去により 3 人とした］は後に，次に示す高名な学者達と結婚した。その様子は，長女「せき」は呉 黄石，二女「つね」は箕作秋坪，そして三女「ちま」は箕作省吾であった（緒方，1978）。とくに，この関係者の貢献を見ると，呉 黄石（1811-1879）は西洋医学者として（木村，1994），箕作秋坪（しゅうへい）（1825-1886）は幕末の洋学者として大活躍をおこなった（佐藤，1985）。また，箕作省吾の一子の箕作麟祥（りんしょう）（1846-1897）は阮甫の初孫で明治期の法律学者・法学博士として活躍（佐藤，1985; 木村，1994），箕作秋坪の長男の箕作奎吾（1852-1871）は明治新政府の大学校に勤めた（木村，1994）。箕作秋坪の三男の箕作佳吉（かきち）（1857-1909）が日本人として初めて東京帝国大学動物学教授になり（鈴木，1985; 木村，1994），さらに箕作秋坪の四男であった箕作元八（げんぱち）（1862-1919）が東京帝国大学理学大学を卒業して東京帝国大学文科の西洋史教授となっていた（高峰，1985; 木村，1994）。

箕作阮甫は，1821（文政 4）年の 24 歳のときに，津山藩の藩侍医になった。箕作阮甫が江戸参勤をはじめたのは 1823（文政 6）年で，1827（文政 10）年には津山に帰った。ところが，1831（天保 2）年 3 月 5 日に家族と共に江戸へ移り，蘭学者箕作阮甫の時代となったのである。1839（天保 10）年 6 月 10 日に箕作阮甫は江戸にある幕府の機関に入り，幕府天文台翻訳員になった。このとき，阮甫は 41 歳であった。ここでは多くの蘭学者と一緒で，天文関係や洋書の研究に励むことができた。幕府は 1855（安政 2）年に天文台を「洋学所」とした。阮甫が 56 歳のときであり，この洋学所事務の中心になった。さらに，1856（安政 3）年には阮甫は蕃書調所（ばんしょしらべしょ）の首席教授職になった。このような阮甫の活躍の様子は小山（1978），北（2008）にもよく示されている。

箕作阮甫は幕府の天文台に勤め始めた 1839 年以来 25 年にわたり幕府の機関に関係しながら蘭学者として，また洋書の翻訳者として大活躍をしていた。箕作阮甫の業績は蘭学やオランダ語での学識に有名であった。彼の翻訳としては 99 部，160 余冊に達するという。

阮甫の初期の業績としては医学関係のものからなり，1836 年から 1848 年にかけて刊行された。医学関係以外の阮甫の著作も多種多彩で，地理書や歴史書などに加えて語学書の他，地質，鉱物，物理，天文，兵器，軍器，造船，電信，文芸，詩文など，驚くほど多彩なものであった。歴史書には西欧の歴史への阮甫の興味がよく含まれている。医学関係を除く阮甫の科学・技術に関する著作は次の通りである（菊池，1978）。

「三兵達吉知幾訳本」	兵書
「歩兵使銃動身軌範」	同
「三兵操治正義」	同
「熕砲点放軌範」	砲術書
「軍用火箭考」	同
「水蒸砲説」	水蒸気砲説
「蒸気砲発明説」	同発明記
「海上砲術全書」	砲術書
「水蒸船説略」	造船書
「衣米針印刷伝信通票略解」	電信機説
「消皮説」	（文書なし）
「地殻図説」	地質書
「密涅刺羅義」	鉱物書
「日阿羅義名目」	地質書
「大地マグネチスミュス」	磁石説
「星学」	天文書
「地質辦証」	地質書
「失表題」	度量衡説

上記の多彩な本の後編には箕作阮甫の地質学を中心とした活躍になっているが，この詳細は次の「箕作阮甫の活躍と日本における地質学誕生との関係」で述べることにする。

最後に，箕作阮甫の体調に触れておく。阮甫はすでに述べたように病身の体質を持っていた。それは喘息の持病があったほか，ときどき頭病，悪心，腰痛，その他で悩んでいたという。このような体調の中で箕作阮甫は洋学者とし

て，医学，地質学，地理学，歴史学，兵法・軍器，造船などの自然科学系で大活躍をしていた。ところが，新洋学の新時代に箕作阮甫は生涯を閉じることになってしまった。箕作阮甫は 1863（文久 3）年 6 月 17 日に江戸で没したのである（表 1）。亨年 65 歳であった。

Ⅲ 箕作阮甫の活躍と日本における地質学誕生との関係

箕作阮甫の研究活動のなかで最後の大成果を挙げた「地質学」という新規の地球科学について明らかにしたい。この解説には石山（1978）による「箕作阮甫の地理学」と蘭学資料研究会編（1978）：「箕作阮甫の研究．付録」の中の「箕作阮甫著訳書．2 地質・天文・物理学関係」などの文献を参考にするとともに，津山洋学資料館の資料（津山洋学資料館，1990，2008）も重視した。

箕作阮甫の前半は上に述べたように医学関係の著作を中心にしていたが，後半の江戸在住時代は地質・地理・歴史・兵事などを中心とする著作時代であった。とくに後半時代最後の箕作阮甫は地質学や鉱物学などの自然科学への関心が極めて深かった。例えば，1839（天保 10）年 6 月 10 日に出仕した幕府の天文台で海外地理を研究するようになった。阮甫家の大きい成果になった自然地理学総論に加えて，箕作阮甫の地質学・鉱物学に関する訳稿を中心とする著書があった。それらの主な本として，「密涅剌羅義」（みつえしらぎ），「日阿羅義名目」（にちあぎめいもく），「地殻図説」（ちかくずせつ），「地質辦証」（ちしつべんしょう）などがある。これらの著書について，箕作阮甫の後期を特徴づける主要な地質学書として次に解説したい。

「密涅剌羅義」：ミネラロギー; 片かなまじり文，7 葉 1 冊からなる。「日本記聞　下」に含まれている。1852 年刊行のドイツ人スクードレルの蘭訳本とした鉱物学書である。次に示すものが原本に由来したものである: Schoedler, Fr.: Schoedler’s boek der natuur, alcemeene becimselen. Der physica, astronomie, chemie, mineralogie, geologie, physiologie, botanie en zookogie, Naar de vifde hoogduitsche uitgave, bewerkt door J.J. Altheer. Utrecht, W.F. Dannefelser, 1852. xvii, 836 bln.

本書はドイツのスクードレル原著「自然の本」の蘭訳で，物理学，天文学，化学，地質学・鉱物学，植物学，動物学，索引と言う構成からなる。この原著は箕作阮甫が 1854（安政　元）年に長崎出張の帰途に島津斉彬より借りたという。

「日阿羅義名目」：ゲオロギーみょうもく; 片かなまじり文，21 葉 1 冊からなる。ゲオロギーは geologie（地質学）のこと。Fr. Schoedler の「スタードレン金石論云」と K.C. von Leonhard の「ホンレオナルド地質説云」の二編の原著で構成。「ホンレオナルド地質説云」は次の通りである: Leonhard, K.C. von.: Geologie, of natuurlijke geschiedenis der aarde; op algemeen bevattelijke wijze voorgesteld. Door K.C. von Leonhard. Uit het hoogduitsch. Amsterdam, G.J.A. Beijerinck, 1845~47, 3 bln. 23 × 14cm.

「地殻図説」：ちかくずせつ; 平かなまじり文で，34 葉 1 冊からなる。Geologie を地質学とした。地殻断面図の説明で，地殻生成，地殻の成分，化石，沈澱および生物に関係する岩石，地層と初層，第二層，第三層などを述べる。

「地質辦証」：ちしつべんしょう; 平かなまじり文で，88 葉 3 巻 1 冊からなる（第 2 巻が欠けている）。地質弁證とも書く。1861（文久　元）年に出版。地質学の総論（ゲオロギー，ゲオダノシー），第一大地流動の時期，第二火鼓鑄の時期，第三火煙山の時期，第四史伝の時期，などの地質概論と地史学が扱われている。特に，第 1 ページの表題として「第 1 篇　ゲオロギー．名義　地質学」を示している。

この本を出したのは箕作阮甫 63 歳の最晩年であった。こうして彼の自然に関する研究は最高になった。

なお，木村（1994）は，上に紹介した地質関係 4 編（「密涅剌羅義」，「日阿羅義名目」，「地殻図説」，「地質辦証」）に加えて地球磁気学書

（大地マグネチスミュス説）と天文学書 1 冊のほか，地理に関する著書として 14 編を紹介している。

これらの著作を基に，石山（1978）は箕作阮甫が日本で初めて「地質学」の用語を用い，西洋地学の内容に立入ったと言っているが，まさにその通りである。ここにあげた箕作阮甫の地質学関係の記述用語の中で，箕作阮甫は日本で初めて「地質学」用語を使ったことになる。このことは，その後の日本における地質学発展のために極めて重要であり，箕作阮甫は日本地質学用語を作った最大の貢献者であったと言える。

Ⅳ あとがき

本文で明確になったように，日本における用語「地質学」を初めて使ったのは箕作阮甫であった。箕作阮甫は 1850 年代のオランダでの用語“geologie”を「地質学」と訳して日本での使用を始めたのである。その後の日本の地質学発展の様子は今井（1966）や矢島（2008）が明らかにしている。しかし，日本の新しい用語「地質学」の形成に今井（1966，1970）は帆足万里の貢献を重視したが（岡田，2008），帆足万里は彼の著書の中で用語「地質学」を使っていなかったことを拙著で示した（岡田，2008）。こうして，本書では日本の用語「地質学」が箕作阮甫によって初めて使用されたことを明確にすることができた。

日本の極めて重要な用語「地質学」ができたことをこの著書になかで明確にし，箕作阮甫の重要な成果を評価することができた。さらに，この前の拙著（岡田，2008）で帆足万里の貢献の様子も明確にしたが，本書出版の機会に，これら二つの拙著の重要性を深く強調したい。

また，幕末期の日本で自然科学の進展に大きい貢献を示した箕作阮甫の成果は岡山県津山市にある現在の「津山洋学資料館」（津山洋学資料館，1990，2008）によく示されているが，ここではさらに箕作阮甫が用語「地質学」の作成に大貢献したことも取り上げて欲しい。

以上のとおり，ここに日本の最初の「地質学」用語の形成を明らかにすることができた。

謝 辞: 本書の執筆でも深い関心を寄せられた故松本達郎先生・故勘米良亀齢先生・首藤次男先生（九州大学名誉教授），唐木田芳文先生（西南学院大学名誉教授），故都城秋穂先生（ニューヨーク州立大学名誉教授），大森昌衛先生（麻布大学名誉教授），中沢圭二先生（京都大学名誉教授），関 陽太郎先生（埼玉大学名誉教授），黒田吉益先生（信州大学名誉教授），橋本光男先生（茨城大学名誉教授），高橋 清先生（長崎大学名誉教授），高柳洋吉先生（東北大学名誉教授），水野篤行先生（元，愛媛大学教授），諏訪兼位先生（名古屋大学名誉教授），松本徰夫先生（山口大学名誉教授），小畠郁生先生（国立科学博物館名誉館員），島津光夫先生（新潟大学名誉教授），松田時彦先生（東京大学名誉教授），志岐常正先生（京都大学名誉教授），水谷伸治郎先生（名古屋大学名誉教授），秋山雅彦先生（元，信州大学教授）に厚くお礼申しあげる。とくに，東北大学の高柳洋吉名誉教授には多数の関係資料をご紹介いただいた。

また，現在の岡山県津山市における箕作阮甫の資料確認についてご教示をいただいた津山市津山洋学資料館下山純正館長と小島 徹学芸員に大変お世話になった。

図 2 箕作阮甫胸像（津山文化センター前の像．鈴木茂之撮影）

V 引用文献

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表 1 箕作阮甫の生涯

1799（寛政 11）	年令 1	9 月 7 日に津山の西新町で誕生.
1809（文化 6）	11	永田桐隠に漢籍を学ぶ.
1810（文化 7）	12	兄亡くなり，12 月 28 日に自分自身を「阮甫」と改名.
1813（文化 10）	15	小島天楽に漢籍を学ぶ.
1815（文化 12）	17	京都に出て漢方医学を学ぶ.
1819（文政 2）	21	京都留学より津山に帰る.
		11 月に津山藩大村成美の養女「とい」（登井）と結婚.
1822（文政 5）	24	津山藩の侍医となる．さらに江戸に出る.
1823（文政 6）	25	5 月 20 日長女「せき」誕生.
1826（文政 9）	28	江戸幕府のシーボルトと会見.
1827（文政 10）	29	津山に帰る.
1828（文政 11）	30	6 月 16 日次女「つね」誕生.
1830（天保 元）	32	11 月 8 日に江戸 10 ヵ年を命じられる.
1831（天保 2）	33	3 月 15 日に江戸に移る.
1832（天保 3）	34	8 月 20 日三女「ちま」誕生.
1834（天保 5）	36	「医療正始」第 1 編出版.
1839（天保 10）	41	6 月 10 日天文台出仕，蛮書和解方となる.
1842（天保 13）	44	「和蘭文典」出版.
1843（天保 14）	45	「海上砲術全書」出版.
1844（弘化 元）	46	「新製輿地全図」出版.
1846（弘化 3）	48	7 月 1 日江戸定府を命じられる.
1847（弘化 4）	49	緒方洪庵の門に入る.
1848（嘉永 元）	50	「水蒸船説略」の訳を始める.
1853（嘉永 6）	55	長崎に行く.
1854（安政 元）	56	長崎より 2 月 24 日江戸に帰る.
		「海国図誌」を出版.
1855（安政 2）	57	3 月に隠居.
1856（安政 3）	58	4 月に蕃書調所教授職となる.
1861（文久 元）	63	「地質辧証」を出版.
1862（文久 2）	64	12 月 28 日に幕府に召出される.
1863（文久 3）	65	6 月 16 日江戸で歿す.

OKAYAMA University
Earth Science Reports
Vol. 16, No. 1, 9-28, (2009)

Sr–Nd isotopic compositions of Paleopr oterozoic metavolcanic rocks from the southern Ashanti volcanic belt, Ghana

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Neodymium (Nd) and strontium (Sr) isotopic data are presented for Paleoproterozoic metavolcanic rocks in the southern part of the Ashanti volcanic belt of Ghana. The metavolcanic rocks are predominantly basalts/basaltic andesites and andesites with minor dacites. Two types of basalts/basaltic andesites (B/A), Type I and Type II, have been identified. The Type I B/A are stratigraphically overlain by the Type II B/A, followed by the andesites and the dacites. The analyzed volcanic rocks commonly have low initial $^{87}Sr/^{86}Sr$ ratios consistent with previous studies on Paleoproterozoic rocks from the West African craton. The LREE-depleted, tholeiitic Type I B/A exhibit back-arc basin geochemical signatures and show high positive epsilon Nd (i.e., ε_{Nd} (2.1 Ga) = +3.89 to +7.21), which suggest a long term depleted source and also indicate that they were produced in an entirely oceanic environment devoid of influence of continental crust. The isotope signatures are thus consistent with the previously published trace element data of the Type I basalts/basaltic andesites in suggesting that their parent magma was generated from a depleted mantle. The Type I B/A have Nd model ages (T_{DM2}) of 1.83–2.09 Ga similar to their formation ages, suggesting that they were juvenile at their time of formation. The andesites and the Type II B/A andesites show LREE-enriched patterns and exhibit characteristics of subduction zone-related magmas, and show initial ε_{Nd} (2.1Ga) values of –1.15 to + 1.35 and Nd model ages (T_{DM2}) of 2.32–2.58 Ga. The LREE-enriched dacitic porphyry also exhibits characteristics of subduction zone-related magmas, and have initial ε_{Nd} (2.1Ga) value of –2.24 and Nd model ages (T_{DM2}) of 2.64 Ga. The Nd isotopic data confirms the juvenile character of the Birimian crust, but also suggests some contributions of a pre-Birimian crustal material (or Archean?) in the genesis of some of the metavolcanic rocks. Our isotopic result is consistent with the island arc complex model which views Paleoproteozoic terranes of West Africa in the context of subduction–accretion processes.

Keywords: Sr–Nd isotopes, petrogenesis, tectonic setting, Birimian metavolcanics, Ashanti volcanic belt

I. Introduction

The processes involved in crustal evolution/growth of continents over geologic time and the nature of the material extracted from the mantle to produce the protocrust, especially during the Achean and Proterozoic, have been surrounded with controversies. While some believe that continental crust grows through time by accretion of island arcs, others are of the view that crustal growth may have resulted from remelting of sinking mafic bodies in a geodynamic context devoid of plate tectonics, underplating of rift/ ocean island magmas or eruptions of continental flood basalts (references in Abouchami et al., 1990). The Archean–Proterozoic boundary (which is arbitrarily put at 2.5 Ga) is considered as a major marker in the changes of the style of crustal evolution in the earth's history. Researches carried out in the greenstone belts of Precambrian shields of Canada, the southwestern United States, western Australia, Greenland, Scandinavia, southern Africa and southern India tend to agree with this major change (cf. Sylvester and Attoh, 1992), thereby supporting the assertion of a worldwide, fundamental change in crust-mantle evolution at the Archean–Proterozoic boundary.

Studies of the Paleoproterozoic 'Birimian' rocks of the Leo-Man shield of the West African Craton (WAC;

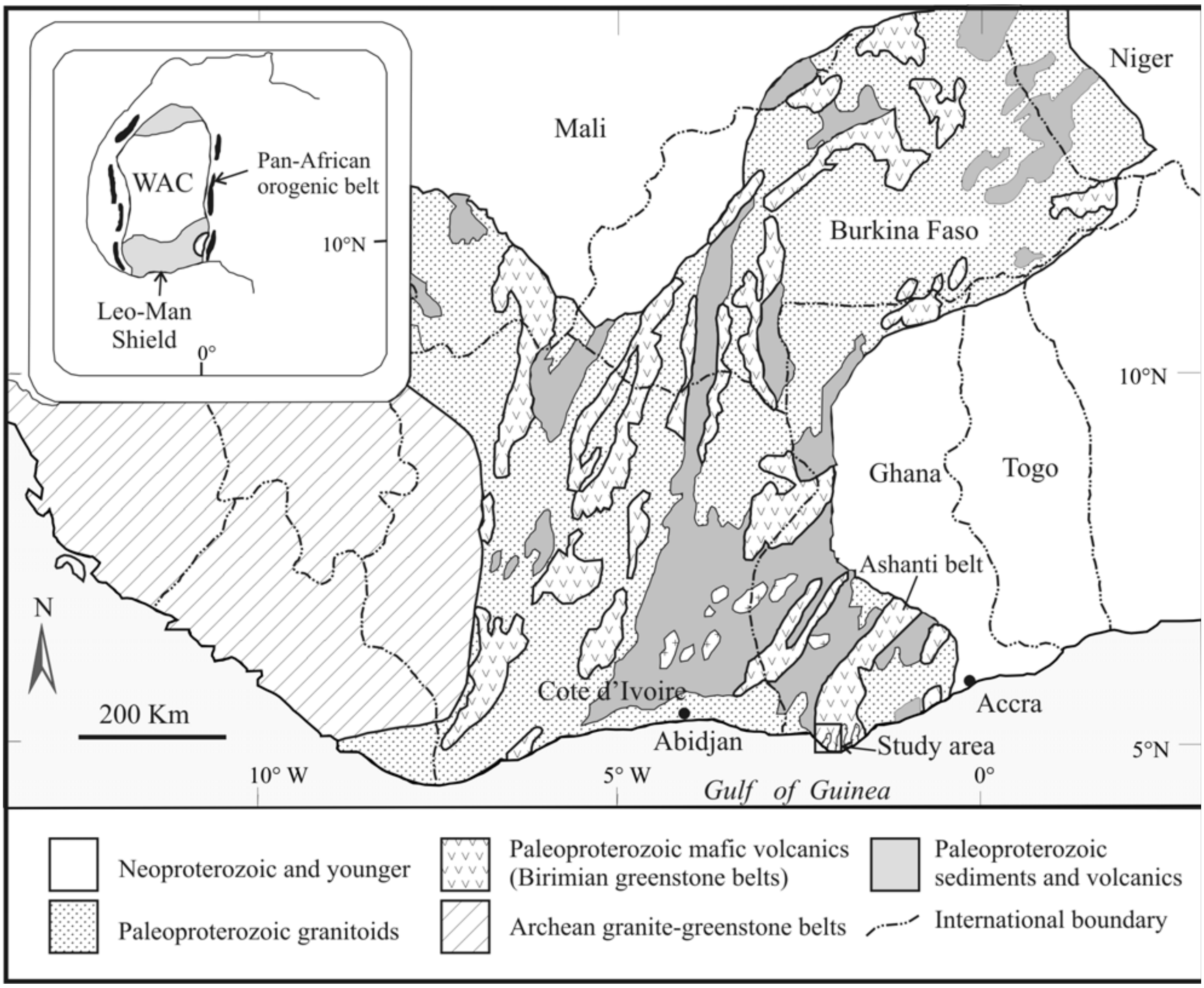

Fig. 1. Geologic sketch of the Leo-Man shield of the West Africa craton (WAC; inset) showing the Paleoproterozoic (Birimian) greenstone belts and the location of the study area (adopted after Attoh et al., 2006).

Fig. 1) have, however, yielded interesting and provocative findings for reconsideration of models regarding the geodynamic evolution of greenstone belts of Precambrian shields. For instance, the ~2.1 Ga Birimian magmatism in the West Africa craton took place between 2.4 and 1.9 Ga, thus bridging a major gap in mantle activity and crustal evolution for a period thought to be quiescent in the North America–Europe regions (Nelson and DePaolo, 1985; Patchett and Arndt, 1986) and Australia (McCulloch, 1987). It has also been suggested from geochemical studies of the Birimian volcanic rocks (Sylvester and Attoh, 1992) and the metasedimentary rocks (Asiedu et al., 2004) from Ghana that the Birimian terrane of Ghana and probably the West African craton may not conform to the worldwide fundamental change in the crust–mantle evolution at the Archean–Proterozoic boundary. Understanding the Birimian magmatism in the West African craton could unequivocally be useful in the appraisal of the crustal growth of continents in the earth's evolution history.

The main Paleoproterozoic crustal growth event (or Eburnean orogeny) in the WAC represents a major juvenile crust-forming process with involvement of only a negligible Archean crustal components (e.g., Abouchami et al., 1990; Liégeois et al., 1991; Boher et al., 1992; Taylor et al., 1992; Pawlig et al., 2006). However, the context of crustal accretion and tectonic setting as well as the lithostratigraphic feature of the Paleoproterozoic rocks has been a subject of contrasted

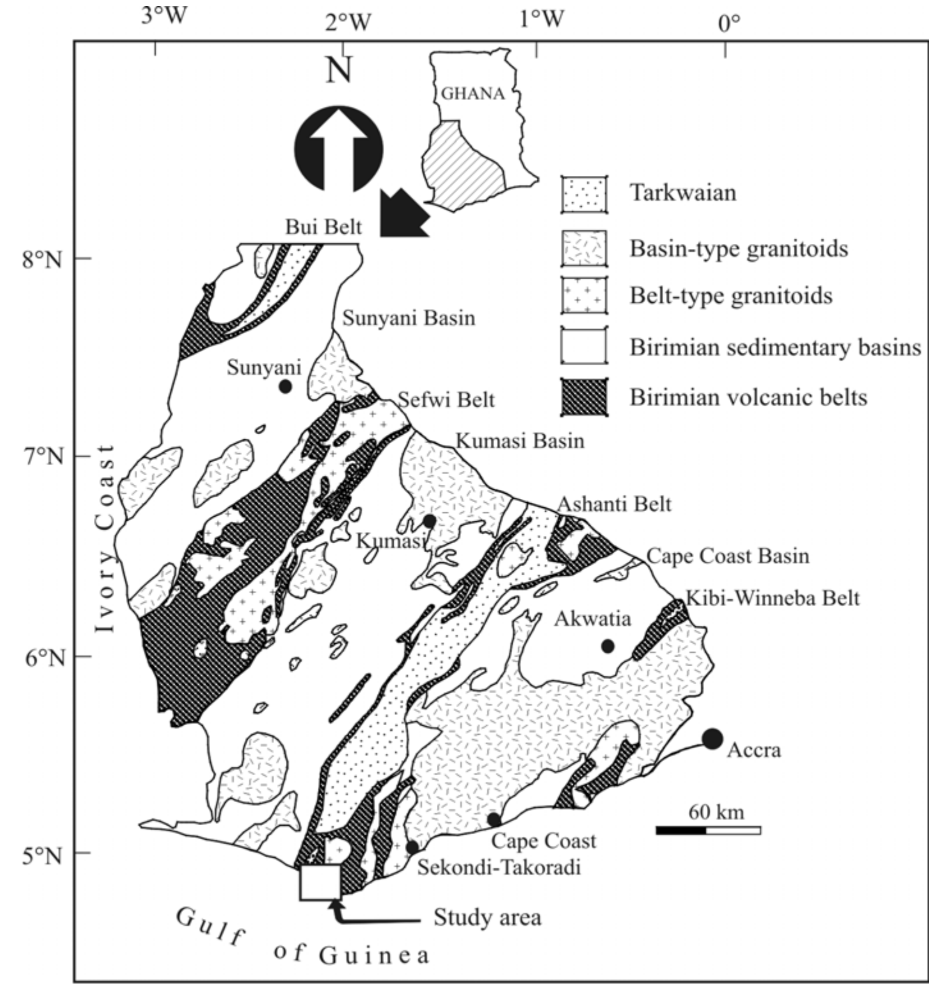

Fig. 2. Simplified geological map of Ghana showing the study area.

interpretations. The lithostratigraphic succession of these rocks (i.e., the volcano-sedimentary and bimodal volcanics) has over the years been contentious with respect to whether the sedimentary unit lies below (e.g., Junner, 1940; Milési et al., 1992; Feybesse and Milési, 1994), or above (e.g., Tagini, 1971; Hirdes et al., 1996; Asiedu et al., 2004) the volcanic unit. It is now widely accepted that the two units formed quasi-contemporaneously, as lateral facies equivalents (e.g., Leube et al., 1990). In terms of the tectonic setting in which the rocks were formed, interpretations vary between generation of the Birimian juvenile crust in arc environments (e.g., Sylvester and Attoh, 1992; Mortimer, 1992; Pohl and Carlson, 1993; Asiedu et al., 2004; Dampare et al., 2005) and from plume-related magmatism (e.g., Abouchami et al., 1990; Lompo, 2009). Two main models have been proposed to account for the tectonic processes involved in the formation of the rocks: accretionary orogeny (e.g., Abouchami et al., 1990; Boher et al., 1992; Davis et al., 1994; Feybesse and Milési, 1994; Hirdes et al., 1996; Ledru et al., 1994; Hirdes and Davis, 2002) versus transcurrent tectonics models (e.g., Bassot, 1987; Doumbia et al., 1998; Pouclet et al., 1996).

Avalaible $^{207}Pb/^{206}Pb$ and U–Pb zircon geochronological data on the granitoids intruding the Birimian terrane tend to suggest that Paleoproterozoic crust formation in the West African craton are marked by two successive major episodes of coeval plutonism and volcanism (i.e., Birimian *sensu stricto* followed by Bandamian). A pre-Birimian crustal growth episode has also been suggested by some workers (e.g., Gasquet et al., 2003; Pawlig et al., 2006).

The Paleoproterozoic terrane of Ghana is mainly characterized by northeast–southwest trending volcanic belts with intervening basins, and both the belts and basins are intruded by granitoids of Proterozoic age (Fig. 2). The southern part of the Ashanti greenstone belt, one of the volcanic belts of Ghana, is characterized by three volcanic lobes, made up of basaltic flows, andesitic lavas, pyroclastic and sedimentary rocks, with granite-diorite plutonic suites occupying intervening positions. Mafic–ultramafic plutonic rocks also occur in this part of the volcanic belt. Some petrographic and geochemical data are available on these rocks (e.g., Sylvester and Attoh (1992; Loh and Hirdes,1999; Attoh et al., 2006; Dampare et al., 2005; Dampare, 2008; Dampare et al., 2008; Dampare et al., in review). However, isotopic data are for the metavolcanic, granitoids and the mafic–ultramafic rocks. Isotopic studies have mostly been restricted to the granitoids of which U–Pb zircon ages (Taylor et al., 1988, 1992; Hirdes et al., 1992; Opare-Addo, 1993; Loh and Hirdes, 1999; Attoh et al., 2006) and Sr–Nd data (Taylor et al., 1988, 1992) are available. Loh and Hirdes (1999) also determined U–Pb zircon ages of a felsic pyroclastic unit within the volcanic pile.

The objectives of this paper are: (1) to present chemical data of Paleoproterozoic metavolcanic rocks exposed in the southern Ashanti greenstone terrane in southwestern Ghana; and (2) to discuss the petrogenesis and the tectonic setting of the metavolcanic rocks.

II. Geological setting

Paleoproterozoic rocks, which comprise the bulk of the Birimian terrane of the Leo-Man shield (Fig. 1), form a significant portion of the West African Craton (WAC) to the east and north of the Archean Liberian cratonic nucleus. The Paleoproterozoic terrane is characterized by narrow sedimentary basins and linear, arcuate volcanic belts and intruded by several generations of granitoids (Leube et al., 1990; Hirdes et al., 1996; Doumbia et al., 1998), and corresponds to a period of accretion during the 2.1–2.0 Ga Eburnean orogeny (e.g., Abouchami et al., 1990). The Eburnean tectono-thermal event was accompanied by the deformation and the emplacement of syn- to post-orogenic tonalite–trondhjemite–granodiorite (TTG) and granite plutons along fractures (e.g., Leube et al., 1990; Vidal and Alric, 1994; Feybesse et al., 2006). The event resulted in widespread metamorphism mostly under greenschist-facies conditions (e.g., Oberthür et al., 1998; Béziat et al., 2000). Also, evidence of medium amphibolite–facies metamorphism may be observed in the vicinity of granitoid plutons (e.g., Junner, 1940; Eisenlohr and Hirdes, 1992; Debat et al., 2003).

In Ghana, the Paleoproterozoic supracrustal rocks are subdivided into the Birimian and Tarkwaian. The Birimian rocks comprise a series of subparallel, roughly equal-spaced northeast trending volcanic belts of volcanic rocks separated by an assemblage of sedimentary rocks. The Birimian rocks were traditionally classified into a two-fold lithostratigraphic and chronological system, consisting of an older metasedimentary sequence (referred to as Lower Birimian) and a younger volcanic sequence of predominantly tholeiitic basalt and pyroclastic rocks (referred to as Upper Birimian) (e.g., Junner, 1940). Leube et al. (1990) have indicated that the volcanic and sedimentary rock sequences formed contemporaneously as lateral facies equivalents. The Birimian rocks are overlain by the detrital Tarkwaian sedimentary rocks in almost all the prominent volcanic belts, and both formations were subjected to similar deformation events, which involved compression along a southeast–northwest trending axis (Eisenlohr and Hirdes, 1992; Blenkinsop et al., 1994). The Tarkwaian rocks are made up of conglomerates, sandstones and subordinate shale, with much of the clastic material in the sediments derived from the adjacent Birimian units (Tunk et al., 2004). Two main suites of granitoids, namely the Dixcove- (or belt-type) and Cape Coast-type (or basin-type) granitoids, intrude the Birimian terrane of Ghana. The metaluminous, relatively Na-rich, Dixcove-type granitoids, which are dominantly hornblende- to biotite-bearing granodiorite to diorite, monzonite and syenite, intrude the Birimian volcanic belts and they may be coeval with the volcanic rocks (Eisenlohr and Hirdes, 1992). The Cape Coast-type granitoids, which are predominantly peraluminous, two-mica granodiorites, with lesser hornblende- and biotite-bearing granodiorites, are emplaced within the Birimian sedimentary basins. The Dixcove granitoids have been dated at 2172 ± 1.4 Ma in the Ashanti Belt (Hirdes et al., 1992) and the Cape Coast type between 2104 ± 2 and 2123 ± 3 Ma in the Kumasi Basin (Oberthür et al., 1998). Other types of granitoids include the localized, biotite and hornblende Winneba granitoids, and the K-rich Bongo granitoids which cut the Tarkwaian rocks in the volcanic belts. The Cape Coast-, Kumasi-, Dixcove- and Bongo-type granitoids have strong mantle affinities whereas the Winneba-type has an Archean sialic precursor (Sm–Nd model age of ~2.6 Ga) (Taylor et al., 1992). The Winneba-type granitoids are believed to be the only rock suite in Ghana which show evidence for a significant magmatic contribution from an Archean continental crust in their genesis.

Tectonic evolution models proposed for the Paleoproterozoic rocks of Ghana have commonly considered two deformation events. In the first event, the Birimian rocks were deformed and intruded by granitoid, and was later uplifted and eroded. The erosional products deposited in a series of grabens located within the volcanic belts to form the Tarkwaian. During the second event, which involved renewed folding (cf. Moon and Mason, 1967; Ledru et al., 1988) or gravity tectonic processes (Leube et al., 1990), both the Birimian and Tarkwaian were deformed. Recently, however, it has been suggested that both the Birimian and Tarkwaian were subjected to a single, progressive deformation event, which involved compression along a southeast-northwest trending axis, resulting in folding and thrusting with subsequent flattening and localized oblique-slip shearing (e.g., Eisenlohr, 1992; Eisenlohr and Hirdes 1992; Blenkinsop et al., 1994; Allibone et al., 2002). Also, the tectonic setting in which the Paleoproterozoic rocks of Ghana formed is contentious, and interpretations of tectonic models for the Birimian terrane of Ghana basically varies between the intracratonic rift setting and island arc complexes. Leube et al. (1990) have proposed an intracontinental rift model for the Paleoproterozoic rocks, whereas arc or subduction related models are preferred by some other workers (Sylvester and Attoh, 1992; Pohl and Carlson, 1993; Loh and Hirdes, 1999; Asiedu et al., 2004; Dampare et al., 2005; Attoh et al., 2006; Dampare et al. 2008). Feybesse et al. (2006) developed a metallogenesis model for the occurrence of gold deposits in Ghana which invoked a combination of continental margin, juvenile magmatism and convergence and collision between an old continent and a juvenile crust for the tectonic environments in which the Birimian greenstone belts were generated. Thus, the Ashanti volcanic belt probably marks the boundary between an Archean continental domain and a Birimian oceanic domain. For Harcouet et al. (2007), the basement in the Ashanti area is likely to be of continental type rather than oceanic, as the results of their numerical modeling using thermal parameters such as thermal conductivity values and heat-production rates are more consistent with the continental basement scenario. In terms of crustal growth event, Taylor et al. (1992) have indicated, from geochronological and isotopic data, that the Birimian of Ghana represents a major early Proterozoic magmatic crust-forming event around 2.3–2.0 Ga by differentiation from a slightly depleted mantle source. Thus, the Birimian of Ghana forms part of the major Proterozoic (Eburnean) episode of juvenile crustal accretion which is recognized in West African Craton and dated at 2.2–2.1 by Abouchami et al. (1990). Davis et al. (1994), also following the accretionary model, have proposed that the Birimian sedimentary basins resulted from accretion of arcs and oceanic plateaus now represented by allochthonous Birimian volcanic belts. Feybesse et al. (2006) have suggested that the evolution of the Paleoproterozoic province of Ghana began around 2.35–2.30 Ga with plutonic activity and the deposition of banded iron formation (BIF)-bearing volcanogenic sediments in basins. According to the authors, the Birimian volcanic belts correspond to a major period of accretion of juvenile basic volcanic–plutonic rocks to an Archean continental domain around 2.25–2.17 Ga, whereas the Birimian sedimentary activity occurred at 2.15–2.10 Ga.

The southern portion of the Ashanti volcanic belt forms three branches referred to as the Axim branch, Cape Three Points branch and Butre branch, with three plutons (locally termed as Prince's Town, Dixcove and Ketan pluton) occupying positions between these branches (Loh and Hirdes, 1999). The volcanic branches are composed of basaltic and andesitic lavas, and pyroclastic rocks. The bulk of these andesitic rocks occur in the Axim volcanic branch where they probably occupy a stratigraphically upper position in the volcanic sequence of the southern Ashanti belt (Loh and Hirdes, 1999). Several mafic–ultramafic rocks are associated with the volcanic rocks in the southern Ashanti volcanic belt (Fig. 2).

Economically, gold and manganese occur in the southern Ashanti belt. Some of the gold deposits are structurally-controlled and mostly occur in the transition zones between the volcanic belts and the basin sediments (e.g., Kesse, 1985; Leube et al., 1990; Hirdes et al., 1993; Oberthür et al, 1996). The two most important regional structural controls of the gold deposits in the southern Ashanti belt seem to be the Axim high-strain zone along the western flank of the Ashanti belt, and the sheared granitoid/greenstone settings (Loh and Hirdes, 1999). The gold-bearing quartz-pebble conglomerates of the Tarwaian also contribute to the gold production from the volcanic belt. Studies carried out on gold mineralization in the Ashanti belt have suggested that the gold-bearing fluids are coeval with (peak-greenschist-facies metamorphism (e.g., Mumin and Fleet, 1995; Loh and Hirdes, 1999). It has also been indicated that higher grade mineral assemblages are mostly restricted to the contact aureoles of the belt type granitoids (e.g., Junner, 1935; Leube et al., 1990). Recently, John et al. (1999), on the basis of mineral chemistry, have suggested that the entire Ashanti belt of southeastern Ghana underwent epidote–amphibolite–facies metamorphism (T = 500–650 °C and P = 5–6 kbar) before experiencing retrograde metamorphism under the greenschist-facies conditions.

III. Previous geochronogy and isotopic studies

Isotopic data on Paleoproterozoic rocks from Ghana are generally scarce and the few available geochronological data from southwestern Ghana have been summarized by Pigios et al. (2003). On the basis of Sm–Nd whole-rock isochron and model ages, Taylor et al. (1992) constrained the age of Birimian supracrustal rocks of western Ghana to 2166 ± 66 Ma.

Most radiometric dating on the Birimian of Ghana indicates that volcanic belts were deposited between 2250 and 2186 Ma whereas the sediments were deposited 100–60 Ma later into basins (Leube et al., 1990; Davis et al., 1994; Oberthür et al., 1998; Pigois et al., 2003). Feybesse et al. (2006), however, suggested that the Birimian volcanic belts corresponded to a major period of accretion of juvenile basic volcanic–plutonic rocks to an Archean continental domain around 2.25–2.17 Ga, whereas the Birimian sedimentary activity occurred at 2.15–2.10 Ga. Using SHRIMP II U–Pb analyses of detrital zircons, Pigois et al. (2003) constrained the maximum age of sedimentation of the Tarkwaian, which overlie the Birimian rocks, to 2133 ± 4 Ma. The Birimian and Tarkwaian rocks host the major gold deposits in Ghana, with most of the gold deposits concentrated along the western flank of the Ashanti belt. The major epigenetic lode-gold event occurred late in the Eburnean orogeny, after the peak of metamorphism was reached (Oberthür et al., 1998; John et al., 1999; Yao et al., 2001; Pigois et al., 2003). Using hydrothermal rutile, Oberthür et al. (1998) have obtained the age of hydrothermal alteration to be 2092±3 and 2086±4 Ma. Recently, Pigois et al. (2003) have determined the most robust age of 2063 ± 9 Ma for gold mineralization in Ghana from SHRIMP II U–Pb analyses of hydrothermal xenotime.

Previous isotopic works on rocks from the southern Ashanti volcanic belt include the contributions from Taylor et al. (1988, 1992), Hirdes et al. (1992), Opare-Addo (1993), Loh and Hirdes (1999) and Attoh et al. (2006). Most of these studies have been restricted to the granitoids, and the age of the metavolcanic rocks are mainly constrained by the ages of the granitoids. Opare-Addo et al. (1993) reported precise TIMS U–Pb zircon ages of 2172 ± 4 Ma for the migmatites and granitoids from the Ketan pluton, which is located to the eastern margin of the Butre branch. Hirdes et al. (1992) obtained a precise TIMS U–Pb zircon age of 2174 ± 2 Ma for a tonalite from the Dixcove pluton, which outcrops between the Butre and the Cape Three Points volcanic branches. Taylor et al. (1988, 1992) reported initial $^{87}Sr/^{86}Sr$ ratio of 0.7017 ± 8, source $^{238}U/^{204}Pb$ (model) of 7.84 and Sm–Nd model age (T_{DM}) of 2.20–2.24 Ga for the Dixcove pluton, and concluded that the rocks were largely juvenile in character. Attoh et al. (2006) have recently determined TIMS U–Pb zircon ages of 2159 ± 4 Ma for a granodiorite from the Prince's Town pluton, which occupies the intervening position between the Cape Three Points and the Axim branches. The age obtained for the Prince's Town pluton is younger than the ones yielded by the Dixcove and Ketan plutons and consistent with the post-tectonic field relations of the Prince's Town pluton, which typically shows well-preserved, undeformed, magmatic textures (Attoh et al., 2006). The age of the Prince's Town pluton has constrained a minimum age for the Aketakyi ophiolitic complex, as it intrudes the western flank of the complex. Loh and Hirdes (1999) obtained TIMS U–Pb zircon ages of 2266 ± 2 Ma for a felsic pyroclastic unit within the volcanic pile. According to the authors, these high zircon values could probably indicate the presence of early Birimian volcanic activity. Dampare et al. (in review) have recently carried out Sr–Nd isotopic studies of the ultramafic and mafic rocks as well as the Prince's Town pluton, and they have provided Nd isotopic evidence for a possible contamination of the juvenile Birimian crust of the southern Ashanti belt by a significant amount of a pre-Birimian crustal material (or Archean?).

IV. Field relations and petrography

A detailed description of the field relations among the various rocks of the southern Ashanti volcanic belt is provided by Loh and Hirdes (1999). The volcanic rocks, mainly composed of basaltic and andesitic lavas, and pyroclastic rocks, form three major NE–SW trending lobes/branches referred to from east to west as the Butre branch, Cape Three Points branch and the Axim branch with three granitoid intrusives (locally called the Prince's Town, Dixcove and Ketan pluton) occurring at intervening positions between them (Fig. 3). The Butre branch consists predominantly of massive, pillow basalt flows; the Cape Three Points branch is characterized by interlayered pyroclastic rocks, basaltic flows and small subvolcanic intrusions, with pyroclastics being more prominent than flows whereas the Axim branch consists of interlayered volcanics and pyroclastic rocks in roughly even proportions, with more than 50% of the volcanics being of andesitic composition (Loh and Hirdes, 1999). The lithostratigraphy of the southern Ashanti volcanic belt is poorly established. According to Loh and Hirdes (1999), the volcanic rocks of the Butre branch form the bottom part of the total volcanic pile, followed by that of the Cape Three Points branch, and the volcanic rocks of the Axim branch occupy the upper stratigraphic position. There is, however, a problem with this stratigraphy, as it cannot explain the overall synclinal structure of the Ashanti volcanic belt (Dampare et al., 2008). For Sylvester and Attoh (1992), the volcanic pile of the Dixcove belt (same as the Cape Three Points volcanic branch) consists of 500m of pillow breccias and hyaloclastites at the base, 3000m of pillowed, massive lava flows in the middle, and 1000m of aquagene tuffs interbedded with massive flows at the top, capped by a manganese-oxide and manganese-carbonate deposit, but this has been questioned by Loh and Hirdes (1999).

Ultramafic to mafic complexes as well as isolated mafic bodies are associated with this greenstone–granitoid belt. The ultramafic–mafic rocks occur as two discrete bodies, i.e., Aketakyi ultramafic complex (UMC) and Ahama ultramafic body. The Ahama ultramafic body was first discovered and mapped by Loh and Hirdes (1999). It is located in the area of the headwaters of the Ahama stream in the Axim volcanic branch. The rocks are poorly exposed. The Ahama rocks are predominantly pyroxenites. The Aketakyi UMC, which was previously mapped by Bartholomew (1961) and Loh and Hirdes (1999), has recently been studied by Attoh et al. (2006). The rocks are well exposed on the coast and form a jagged coastline from Aketakyi to Cape Three Points. The western contact of the Aketakyi UMC is marked by a fault and considered to represent the lower section of the complex, where it is made up of ~2.0 km thick zone of coarse-grained peridotites, including harzburgite and dunite (Attoh et al., 2006). The rocks are variably serpentinized and altered to various degrees. Layering in the ultramafic complex appears to follow the regional northeast strike of the volcanic belt. This layering is more pronounced in the gabbroic rocks, which occur in the eastern contact zone. Rodingite has been identified in the ultramafic rocks. The mafic bodies include the Axim gabbro, Kegyina gabbro and gabbroic rocks from the Aketakyi area. Although the gabbroic rocks have not been dated, they appear to have different origins as well as emplacement records. According to Loh and Hirdes (1999), most of the gabbroic rocks appear to predate the Eburnean tectono-thermal event. The Axim gabbro is associated with argillite and andesitic lavas in the transition zone between the Kumasi basin and the Ashanti belt. The rocks are massive and show various degrees of alteration. The Kegyina gabbro is poorly exposed and deeply weathered. It is located within the Prince's Town pluton toward its margin, and probably represents the mafic boarder of the Prince's Town pluton (Loh and Hirdes,

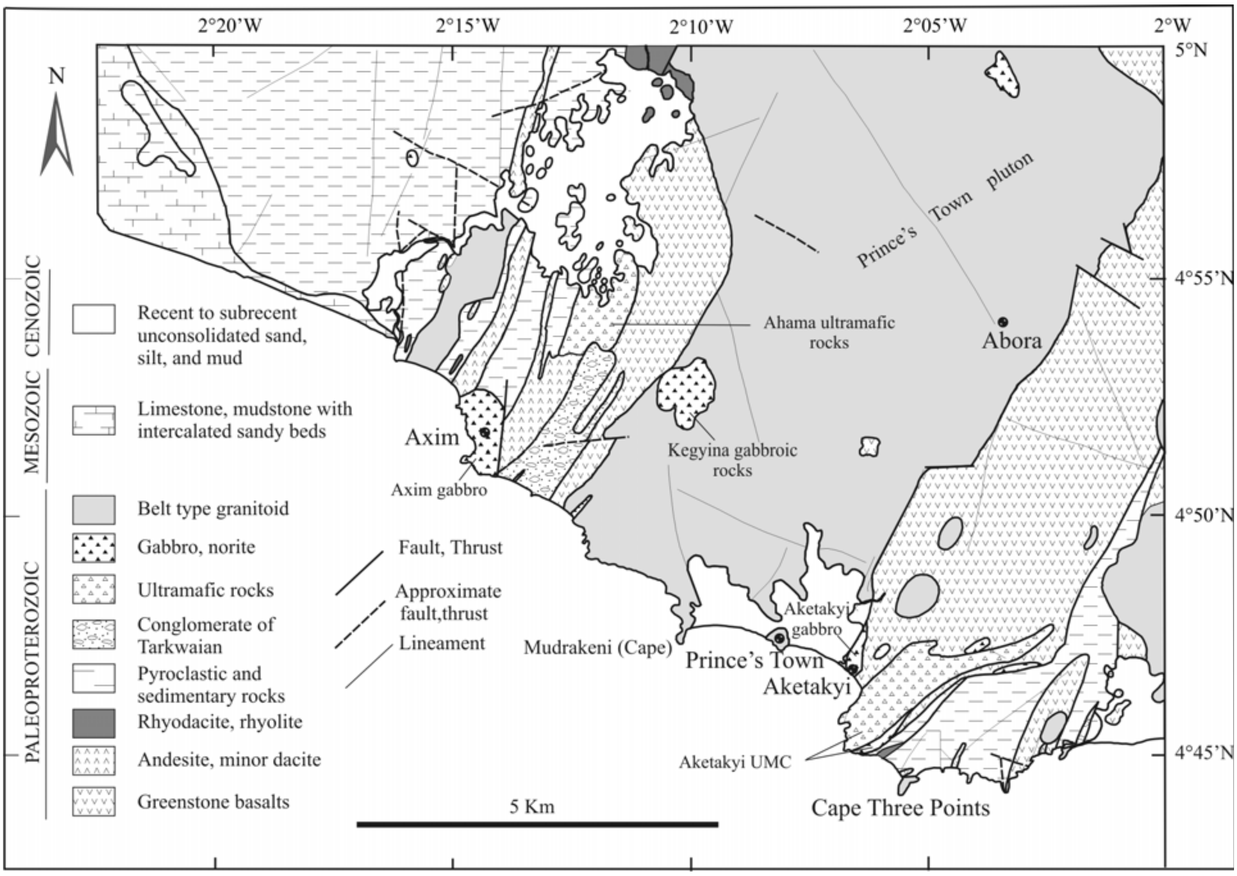

Fig. 3. Geological map of the southern Ashanti volcanic belt of Ghana showing the Axim (left) and Cape Three Points (right) volcanic branches (after Loh and Hirdes, 1999).

1999). The rocks are coarse-grained and massive. The gabbroic body, which intrudes the area north of Aketakyi town, is considered as an equivalent of the Kegyina gabbro. This body is made up of fine- to coarse-grained types. The Aketakyi gabbroic rocks are associated with the Aketakyi ultramafic complex, and they commonly occur as strongly foliated rocks at the outer flanks of the ultramafic complex. The ultrabasic nature of some analyzed gabbroic rocks led Loh and Hirdes (1999) to suggest that the gabbroic rocks could be part of the main Aketakyi ultramafic complex, and Attoh et al. (2006) have also indicated that these rocks represent the upper section of the ultramafic complex. The gabbroic zone of the Aketakyi ultramafic complex is stratigraphically overlain by a sheeted-dyke complex, which consists of basaltic hyaloclastites intercalated with fine-grained sediments including chert and disrupted basaltic-andesitic sills. The ultramafic complex with the overlying sheeted-dyke complex and associated volcanics has been suggested to represent a cross-section of a Paleoproterozoic oceanic crust (Attoh et al. 2006).

Metavolcanic rocks, mainly andesitic and basaltic lavas flows, were collected from the Axim and the Cape Three Points volcanic branches of the southern Ashanti greenstone belt for this study (Fig. 3). The analyzed basalts/basaltic andesites are dark green to dark grey, fine-grained and massive. They are mostly aphyric and sparsely porphyritic, with less than 10% phenocrysts set in a microcrystalline groundmass. The primary minerals of the basalts/basaltic andesites comprise mainly plagioclase and pyroxene which are either partially or completely replaced by actinolite and epidote. Plagioclase which occurs as euhedral to subhedral phenocrysts is usually replaced by epidote and sericite, and in some cases by albite and calcite. Relict pyroxene is observed in some of the specimens. The groundmass consists of lath-like and microlite plagioclase, and secondary quartz, carbonate, chlorite and sericite. Apatite, magnetite and ilmenite occur as the main accessory minerals. Leucoxene may occur as an alteration product of ilmenite. In the foliated types, some whitish bands made up of mainly quartz, calcite and albite alternate with dark greenish bands consisting of chlorite, epidote and actinolite.

The andesites and some dacites were collected from the Axim branch. They are greenish to dark grey, fine- to medium-grained and massive. The andesites are mostly microporphyritic. Plagioclase constitutes the main phenocryst phase in the rocks. Some of the phenocrysts might be originally a mafic mineral which is now replaced by actinolite, chlorite, epidote and sometimes carbonate. Euhedral to subhedral plagioclase crystals are commonly replaced by epidote. The fine-grained groundmass consists mainly of plagioclase, and secondary actinolite, epidote and quartz. Accessory minerals include opaques and apatite, with the opaques sometimes altering to leucoxene. The dacites show a porphyritic texture with few phenocrysts of plagioclase and quartz set in a fine-grained cryptocrystalline matrix, which consists of secondary sericite, epidote, chlorite, as well as accessory pyrite, ilmenite and apatite.

V. Analytical methods

Analysis of major elements of the metavolcanic rocks was carried out on fused discs by the automated X-ray fluorescence spectrometer, Philips model PW 1480, at the Department of Earth Sciences, Okayama University. The trace elements analyses of selected samples were performed at the Activation Laboratories Ltd. (Actlabs), Ontario, Canada, using fusion inductively coupled plasma and inductively coupled plasma-mass spectrometry (ICP-MS). Details of the analytical procedures (i.e., accuracy, precision and standards) were presented by Dampare et al. (2008).

Sr–Nd isotope analyses were conducted at the Center of Instrumental Analysis, Okayama University, Japan. Approximately 70 mg of whole-rock powdered samples were dissolved in a mixture of purified HNO_3–HF–$HClO_4$ in Teflon vials on a hot plate (100–140 °C). Separation of Sr and Nd was carried out using a standard two-column ion-exchange technique, described in detail by Dampare (2008). Isotopic analyses were carried out using a five-collector Finnigan MAT-262 mass spectrometer. Sr and Nd were loaded in 2% H_3PO_4 (repeatedly purified using a cation exchange column) on double Ta-Re filaments and analyzed in the static mode. In order to correct for mass fractionation, the $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ isotopic data were normalized against the values of $^{86}Sr/^{88}Sr = 0.1194$ and $^{146}Nd/^{144}Nd = 0.7219$, respectively. The mean value of the measured $^{87}Sr/^{86}Sr$ ratios of NBS 987 standard was 0.710295±11 (2σm; n=13 analyses). The mean of the measured $^{143}Nd/^{144}Nd$ ratios of JNdi-1 standard gave 0.5121428±9 (2σm; n=5 analyses), which corresponds to a La Jolla Nd standard value of 0.511885.

VI. Geochemistry

Detailed chemical compositions of the southern Ashanti belt metavolcanic rocks have been discussed elsewhere (Dampare et al., 2008), and a summary of these compositions is provided in Table 1. Two types of basalts/basaltic andesites (B/A), Type I and Type II, have been identified with respect to their geochemical features. The Type I B/A, obtained from the Cape Three Points greenstone branch, are characterized by SiO_2

Table 1
Summary of whole-rock major and trace element compositions of metavolcanic rocks from the southern Ashanti volcanic belt

Locality	Cape Three Points Volcanic Branch		Axim Volcanic Branch	
	Type I Basalts/ Basaltic andesites [a]	*Dacitic porphyry* [b]	*Type II Basalts/ Basaltic andesites* [a]	*Andesites and Dacites* [a]
(wt.%)				
SiO_2	47.56?51.76	62.71	48.48–52.87	54.16–68.13
TiO_2	0.62?0.82	0.37	0.54–0.86	0.35–0.77
Al_2O_3	12.57?14.53	14.86	10.41–16.17	13.47–16.14
Fe_2O_3	10.35?12.03	5.61	8.59–9.93	4.32–9.36
MnO	0.17?0.19	0.08	0.13–0.15	0.07–0.18
MgO	6.85?8.16	3.39	3.99–12.67	0.99–5.05
CaO	6.08?10.76	4.91	6.94–9.00	2.95–7.60
Na_2O	0.90?3.19	4.86	2.26–4.51	2.35–5.10
K_2O	0.02?0.31	0.32	0.98–1.22	0.34–2.49
P_2O_5	0.06?0.11	0.09	0.17–0.30	0.09–0.28
Mg#	54.1?60.0	54.5	47.9–71.6	31.1–53.5
(ppm)				
V	224?277	104	168–275	105–202
Cr	670?750	290	250–1060	160–270
Co	31?70	20	35–42	30–48
Ni	220?330	130	100–260	40–120
Cu	40?130	70	30–180	30–70
Zn	70?100	< 30	60	50–90
Rb	2?3	7	17–23	36–70
Sr	75?111	193	249–888	394–756
Y	11.1?15.8	8.8	13.4–16.0	14.1–20.3
Zr	31?41	65	32–53	74–104
Nb	1.0?1.4	1.3	1.4–2.9	3.0–4.2
Cs	0.2?0.3	0.3	0.4–1.3	0.7–5.4
Ba	14?49	95	438–585	456–861
La	1.37?1.91	6.57	3.33–7.11	14.1–24.0
Ce	3.59?5.16	14.8	8.96–18.1	27.8–53.4
Pr	0.54?0.86	1.92	1.27–2.49	3.81–6.68
Nd	2.68?4.98	7.55	5.94–10.5	15.8–26.2
Sm	0.86?1.66	1.76	1.61–2.46	3.64–5.58
Eu	0.416?0.596	1.18	0.621–0.878	1.28–1.68
Gd	1.23?2.35	1.64	1.74–2.45	3.19–4.83
Tb	0.25?0.48	0.25	0.31–0.40	0.50–0.68
Dy	1.78?3.22	1.4	1.90–2.29	2.59–3.60
Ho	0.40?0.67	0.27	0.39–0.45	0.48–0.67
Er	1.27?2.06	0.77	1.19–1.33	1.40–1.97
Tm	0.201?0.324	0.112	0.178–0.199	0.205–0.301
Yb	1.38?2.10	0.77	1.15–1.31	1.32–2.03
Lu	0.225?0.325	0.126	0.177–0.201	0.203–0.316
Hf	0.9?1.2	1.8	1.0–1.4	2.0–2.8
Ta	0.06?0.71	0.1	0.11–1.55	0.21–2.75
Th	0.09?0.15	1.11	0.47–1.08	2.08–3.92
U	0.03?0.23	0.36	0.29–0.46	0.63–1.36

Magnesuim number (Mg#) = 100 x molar $Mg^{2+}/(Mg^{2+}+Fe^{2+})$; [a] From Dampare et al. (2008); [b] This study contents from 47.6 to 51.8 wt.%, MgO contents from 6.85 to 8.16 wt.%, Al_2O_3 contents of 12.57"14.53 wt.%, low TiO_2 contents (0.62"0.82 wt.%), magnesium numbers (Mg#) of 54"60, and high contents of Cr (670"750 ppm), Ni (220"330 ppm), Co (31"70 ppm) and V (224"260). They show flat to slight LREE depletion with $(La/Sm)_N$ = 0.69–1.03, $(La/Yb)_N$ = 0.57–0.71, minor negative and positive Eu anomalies (Eu/Eu* = 0.92–1.24), and slightly negative Nb–Ta anomalies and low Th/Nb ratios (0.06–0.11).

The Type II basalts/basaltic andesites, mainly from the Axim greenstone branch, have SiO_2 contents of 48.5–52.9 wt.%, MgO contents of 3.99–12.67 wt.%, Al_2O_3 contents of 10.41–16.17 wt.%, TiO_2 contents of 0.54–0.86 wt.%, Mg# of 48–72, and Cr (250–1060 ppm), Ni (100–260), Co (35–42) and V (168–275). They possess fractionated REE with $(La/Sm)_N$ = 1.34–2.31, $(La/Yb)_N$ = 2.08–4.25 and minor positive Eu anomalies (Eu/Eu* = 1.09–1.13), and display strong negative Nb anomalies but relatively smaller Ti anomalies and lower Th/Nb ratios (0.35–0.37) compared to the andesites.

The andesites have variable SiO_2 (56.3–63.7 wt.%), MgO (2.98–5.34 wt.%), Al_2O_3 (14.2–16.8 wt.%), TiO_2 (0.49–0.80 wt.%), Cr (160–270), Ni (40–120), Co (30–48), V (105–202) contents, and Mg# (40–54). They show more fractionated REE [$(La/Sm)_N$ = 1.97–2.78; $(La/Yb)_N$ = 4.11–8.48] with minor positive to non-existent Eu anomalies (Eu/Eu* = 0.99–1.15), and display strong negative Nb and Ti anomalies and relatively higher Th/Nb ratios (0.69–95). The dacites have SiO_2 contents of 68.4–69.5%, MgO contents of 1.01–3.39 wt. %, Al_2O_3 contents of 14.9–15.3 wt.%, TiO_2 contents of 0.36–0.46 wt.%, and Mg# of 31–55. The analyzed dacitic porphyry has fairly high Cr (290 ppm) and Ni (130 ppm) contents, and the least contents of Co (20 ppm) and V (104). It also shows fractionated REE [$(La/Sm)_N$ = 2.41; $(La/Yb)_N$ = 6.12] with a high positive Eu anomaly (Eu/Eu* =2.12), and displays strong negative Nb–Ta and Ti anomalies and a relatively high Th/Nb ratio of 0.85 (Table 1).

On the basis of Th–Nb–La–Ce systematics, Dampare et al. (2008) interpreted the negative Th and HFSE anomalies observed in the basaltic and andesitic rocks to reflect a recycled slab-derived lithosphere component instead of crustal contamination. The authors suggested that the metavolcanic rocks were formed in an intra-oceanic island arc-forearc-backarc setting.

VII. Sr–Nd results

Sr–Nd isotopic data including $^{87}Sr/^{86}Sr$ initial ratios, ε_{Nd} (T) and Nd model ages (T_{DM1} and T_{DM2}) are presented in Table 2. The initial $^{87}Sr/^{86}Sr$ ratios and ε_{Nd} values were calculated at an age of 2.1 Ga, which represents the crustal formation time during the Eburnean orogeny (e.g., Abouchami et al., 1990; Boher et al., 1992). The initial isotopic ratios were derived using Rb/Sr and Sm/Nd ratios from the ICP-MS data (Table 1; Dampare et al., 2008). The rocks display low $^{87}Sr/^{86}Sr$ initial ratios (0.697637–0.702423) with the exception of one tholeiitic basalt which has a relatively higher $^{87}Sr/^{86}Sr$ initial ratio (0.706226). The high $^{87}Sr/^{86}Sr$ initial ratio could be due to hydrothermal alteration or a post"emplacement contamination which probably disturbed the Rb–Sr system. The low $^{87}Sr/^{86}Sr$ initial values of D8312A (0.697637) probably suggests alteration of feldspar. The Sr initial ratios values are comparable to those previously reported on the Birimian basaltic and granitic rocks of West Africa (e.g., Abouchami et al., 1990; Boher et al., 1992; Taylor et al., 1992; Gasquet et al., 2003) (Fig. 4). These data show the primitive isotopic Sr signature of the rocks, thereby precluding significant magmatic contributions from Rb-enriched continental crust. The Type I basalts/basaltic andesites show high positive epsilon Nd (i.e., +3.89 to +7.21) isotopic initial ratios at 2.1 Ga. The andesites and Type II basalts/basaltic andesites show a range of initial ε_{Nd} (T) values from –1.15 to +1.35, whereas the dacitic porphyry shows a negative initial ε_{Nd} (T) value of –2.24.

Crustal residence ages (T_{DM}) may be calculated using diverse models of depleted mantle evolution, including the models by DePaolo (1981), Ben Othman et al. (1984), Nelson and DePaolo (1985), Goldstein et al. (1984), McCulloch (1987) and Albarède and Brouxel (1987) with ages varying by several hundreds of Ma. The model of Ben Othman et al. (1984), considered to be representative of the Birimian depleted mantle and used by some workers (e.g., Abouchami et al., 1990; Boher et al., 1992; Gasquet et al., 2003), was adopted for the calculation of one-stage model ages (T_{DM1}) of the rocks. The Nd model age of crustal rocks indicates their time of extraction from the mantle. However, the assumption that a Sm–Nd model age represents an average crustal residence time is valid only if no fractionation of Sm/Nd has taken place since the first separation of its protolith from the mantle source. Since this is not always the case and in order to avoid either underestimation or overestimation of one-stage Nd model ages, a two-stage Nd model age (T_{DM2}) may be calculated for the rocks. The two-stage model age was computed following the procedure of Keto and Jacobson (1987), and using the expression below:

$$T_{DM2} = T_{DM1} - (T_{DM1} - t)\,[(f_{cc} - f_s)/(f_{cc} - f_{DM})],$$

Table 2

Sr-Nd isotopic compositions of Paleoproterozoic metavolcanic rocks from the southern Ashanti volcanic belt, Ghana

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	2s ($X10^{-6}$)	$(^{87}Sr/^{86}Sr)_{2.1}$
Type I B/A						
D9045	<1	75		0.702447	13	
D9047	2	111	0.0521	0.703328	11	0.701751
D9048	2	75	0.0772	0.708562	14	0.706226
D9049	3	103	0.0842	0.70452	13	0.701970
Type II B/A						
D9027	23	249	0.2673	0.708565	10	0.700475
D9028	17	560	0.0878	0.704000	13	0.701342
D9015A	27	888	0.0879	0.704491	12	0.701829
Andesite						
D9033	37	737	0.1452	0.706819	13	0.702423
D8312	70	394	0.5143	0.713205	14	0.697637
Dacitic porphyry						
D90419	7	193	0.1049	0.705347	12	0.702171

Sample	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	2s ($X10^{-6}$)	ε_{Nd} (2.1 Ga)	T_{DM1} (Ga)	T_{DM2} (Ga)	$f_{Sm/Nd}$	$\alpha_{Sm/Nd}$
Type I B/A										
D9045	1.66	4.98	0.2017	0.513002	9	5.79	0.68	1.92	0.025	0.937
D9047	1.40	4.15	0.2041	0.513054	10	6.15	0.23	1.91	0.038	0.943
D9048	1.29	3.90	0.2001	0.513053	7	7.21	0.19	1.83	0.017	0.911
D9049	0.86	2.68	0.1941	0.512801	8	3.89	2.07	2.09	-0.013	0.928
Type II B/A										
D9027	1.61	5.94	0.1640	0.512127	8	-1.15	3.03	2.58	-0.166	0.850
D9028	1.99	8.78	0.1371	0.511824	9	0.19	2.50	2.42	-0.303	0.695
D9015A	2.46	10.5	0.1417	0.511947	7	1.35	2.39	2.31	-0.279	0.705
Andesite										
D9033	5.58	26.2	0.1289	0.511712	7	0.24	2.45	2.41	-0.345	0.653
Dacitic porphyry										
D90419	1.76	7.55	0.1410	0.511754	9	-2.24	2.82	2.64	-0.283	0.744

ε_{Nd} values calculated at 2.1 Ga, relative to the present-day chondritic values of $^{143}Nd/^{144}Nd$ = 0.512638 and $^{147}Sm/^{144}Nd$=0.1967

$f_{Sm/Nd} = [(^{147}Sm/^{144}Nd_{Sample})/(^{147}Sm/^{144}Nd_{CHUR})]-1$

T_{DM1} and T_{DM2} are single-stage and two-stage Nd model ages; T_{DM1} computed according to the depleted mantle model of Ben Othman et al. (1984) and TDM2 following the approach of Keto and Jacobson (1987)

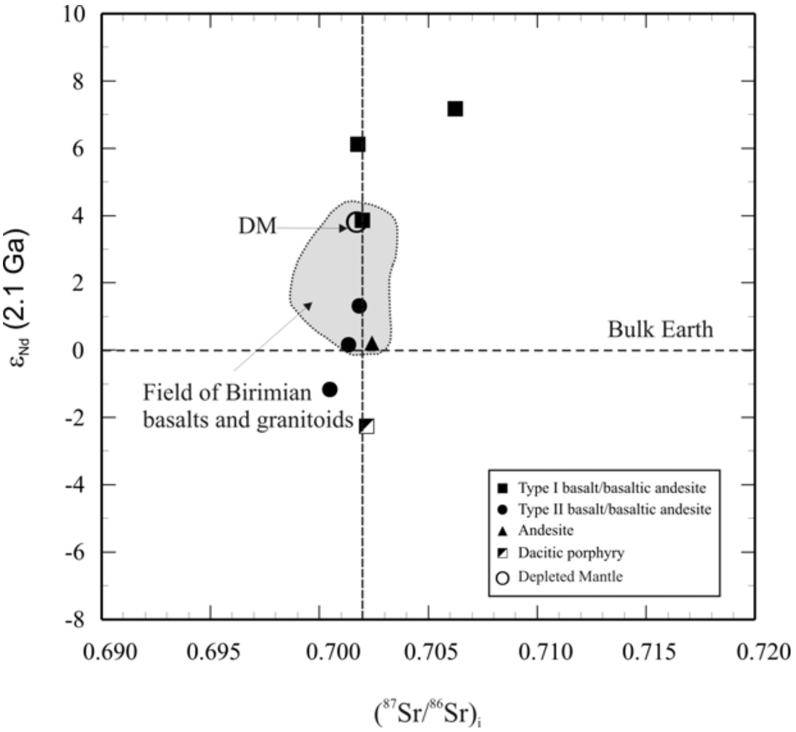

Fig. 4. ε_{Nd} versus initial $^{87}Sr/^{86}Sr$ ratios (calculated at 2.1 Ga) plot for the Paleoproterozoic metavolcanic rocks from the southern Ashanti volcanic belt. Also shown is the field of isotopic composition of Birimian basaltic and felsic rocks (shaded area) from the West African Craton (adopted from Pawling et al., 2006, and the references therein). DM represents contemporaneous depleted mantle from Ben Othman et al. (1984).

where f_{cc}, f_s and f_{DM} are the $f_{Sm/Nd}$ values of the continental crust, the sample and depleted mantle, respectively, and t is the formation age of the rock; and $f_{Sm/Nd}$ is given by the expression:

$$f_{Sm/Nd} = [(^{147}Sm/^{144}Nd_{Sample})/(^{147}Sm/^{144}Nd_{CHUR})] - 1$$

In this calculation, $f_{cc} = -0.4$, $f_{DM} = 0.086426$ and t = 2.1 Ga.

Both one-stage and two-stage model ages have their own uncertainties (references in Dampare et al., 2008). Wu et al. (2005), however, indicated that more accurate results could be obtained for the single-stage model age, if the Sm/Nd fractionation, expressed as $f_{Sm/Nd}$, is limited to the range of –0.2 to –0.6. In this study, the two-stage Nd model age (T_{DM2}) is preferred to the single-stage model age (T_{DM1}) as it displays a more regular pattern. The two-stage Nd model ages (T_{DM2}) calculated for the volcanic rocks are shown in Table 2.

The model ages are 1.83–2.09 Ga for the Type I basalts/basaltic andesites, 2.31–2.58 Ga for the andesites and Type II basalts/basaltic andesites, and 2.64 Ga for the dacitic porphyry.

VIII. Discussions

1. Petrogenesis

The Type I basalts/basaltic andesites (B/A) occupy a lower stratigraphic position of the volcanic pile in the study area, and they are overlain by the Type II basalts/basaltic andesites (B/A), followed by the andesites and dacites. The metavolcanic rocks were derived from a moderately depleted to slightly enriched source. The analyzed volcanic rocks commonly have low initial $^{87}Sr/^{86}Sr$ ratios consistent with previous studies on Paleoproterozoic rocks from the West African Craton. The Type I B/A have Paleoproterozoic Nd model ages (1.83–2.09 Ga; Fig. 5), suggesting that they were juvenile at their time of formation. Their ε_{Nd} (2.1 Ga) values (3.89–7.21) may suggest mantle derived magmas with variable endogenic contamination (i.e., contamination within the mantle rather than the crust, such as inputs from subducted sediments), though to a small extent. The juvenile character of these rocks is confirmed in the plot of ε_{Nd} (2.1 Ga) versus formation age (t) where they either fall in or close to the depleted mantle curve (Fig. 5),

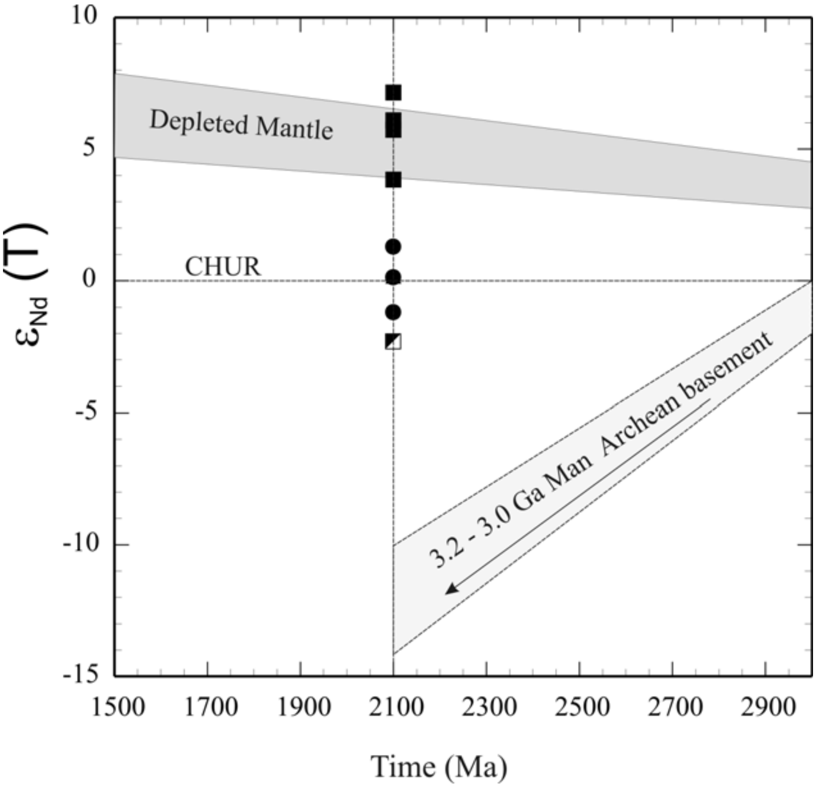

Fig. 5. Plot of ε_{Nd} (2.1 Ga) versus formation age (time) for the Paleoproterozoic metavolcanic rocks from the Ashanti volcanic belt. Data of Archean continental crust is from Kouamelan et al. (1997). Symbols as in Fig. 4.

and are also widely separated from the Nd isotopic evolutionary trend of Archean crust. The Nd isotopic values used for the Archean crust are from Koualeman et al. (1997), who estimated the composition of Archean rocks from Cote d'Ivoire which forms part of the Archean domain of the WAC, since no Archean crust is known in Ghana, at least for now. These Archean rocks have ε_{Nd} (2.1 Ga) values ranging from –10 to –13. As indicated earlier, the ε_{Nd} (2.1 Ga) values of the plutonic rocks range from –5.68 to +4.93, suggesting that the mantle source of the rocks have been contaminated to various degrees by crustal components. Thus, the trace element and isotope signatures of the Type I B/A are consistent in suggesting that their parent magma was generated from a depleted mantle. Leube et al. (1990) have indicated that most Birimian tholeiites of Ghana possess N-MORB chemistry with few showing slight to moderate LREE enrichment. The authors interpret the coexistence of both flat and LREE-enriched patterns of the tholeiites and also the ε_{Nd} value of +2 (Taylor et al., 1988) as indications of slight crustal contamination in the petrogenesis of the tholeiites.

The REE characteristics of the Type II B/A suggest that they could have formed from an enriched source (Dampare et al., 2008). This is consistent with their initial ε_{Nd} (2.1 Ga) values from –1.15 to +1.35, which are less than those of Type I B/A. Their Nd model ages T_{DM2} (2.31–2.58 Ga) are also higher than the formation age (2.1 Ga), which suggest that they may have received some inputs from crustal sources. The trace element data, however, argue against any major crustal contamination (Dampare et al., 2008). The analyzed dacitic porphyry has initial $^{87}Sr/^{86}Sr$ ratio of 0.702171, which indicate minimum crustal contamination of the mantle melt. The ε_{Nd} (2.1 Ga) values of this rock is –2.24, indicative of partly enriched to enriched mantle products. The initial ε_{Nd} value of the rock coupled with geochemical features such as Nb depletion relative to the LREEs, negative Nb–Ta anomalies, relatively high LILE/HREE (or LILE/HFSE) ratios, could point to subduction-related lithospheric mantle sources. Their high negative ε_{Nd} (2.1 Ga) values and Nd model age of 2.64 Ga are possibly indicative of a greater subduction input of an older crustal material. Thus, the Nd model ages of Type II B/A, andesite and dacite are dominated by pre-Birimian (and Neoarchean) ages, and have received variable contributions from pre-Birimian crustal materials, with the most significant contribution recorded in the dacitic porphyry (Fig. 5). The enriched source could have resulted from a recycled slab-derived lithospheric component but not necessarily from an enriched mantle source. It is suggested that the Birimian juvenile crust was contaminated by subduction components, thereby producing heterogeneous mantle beneath the southern Ashanti volcanic belt.

The positive ε_{Nd} values for some of the rocks, including those from LREE-enriched crustal precursors show that their ultimate sources had been previously depleted in Nd relative to Sm when compared to average chondritic Sm/Nd ratios. The very high positive initial ε_{Nd} values (+5.79, +6.15, +7.21) for the Type I B/A indicate that mantle sources with severe time-averaged depletion probably existed in the study area. Until recently when Ngom et al. (2009) obtained an initial ε_{Nd} value of +7.21 for basalt from the Mako volcanic belt of the Kedougou–Kenieba inlier (Senegal), such high initial ε_{Nd} values had not been reported for the Birimian terrane of the West African craton. It can be concluded that those rocks were derived from mantle sources with a time-averaged depletion in Sm/Nd with little or no contribution from significantly older crust. The high positive initial ε_{Nd} values of some of the volcanic rocks strongly suggest that the rocks were produced from the mantle entirely in an oceanic setting (cf. DePaolo, 1988).

The negative ε_{Nd} (2.1 Ga) values of some of the rocks indicate that the rocks cannot be pure mantle-derived juvenile additions to the crust at 2.1 Ga, consistent with the fact that their T_{DM} ages are 200 to 540 Ma older than their emplacement age. The negative ε_{Nd} (2.1 Ga) values can be explained in two possible ways: (1) they represent pure melts of crust that was formed from the mantle at 2300–2640 Ma and their T_{DM} ages faithfully record the mean crustal formation of their protolith, (2) they represent blends of juvenile 2100 Ma crust with an older component, so that the their T_{DM} ages are geologically meaningless. The lack of any documented or known, at least for now, >2300 Ma geological activity in Ghana and the rest of Paleoproterozoic (Birimian) terrane of the West African craton provides an important argument against an interpretation of the rocks as pure melts of 2300–2640 Ma. The second alternative, which is a preferred model, envisages formation of the rocks by mixing of juvenile 2.1 Ga crust material with older material. The lack of correlation between the Sr and Nd isotopes (Fig. 4) makes it difficult to identify the contribution of well-defined source components in the rocks, especially those which have negative ε_{Nd} (2.1 Ga) values.

1.1 Nd modeling: Source characteristics

The ε_{Nd} values can be used to estimate the $^{147}Sm/^{144}Nd$ ratios of the mantle source of the rocks during 2.1 Ga by using the expression (DePaolo, 1988) below:

$$f^{source}_{Sm/Nd} = \frac{\varepsilon_{Nd}}{Q(T_S - T_X)}$$

where T_S = model age of the magma source and T_X = crystallization age of the rocks.
Using T_S = 4.55 Ga where from accretion of the earth started, T_X = 2.1 Ga, Q = 25.09 Ga^{-1} (calculated from present day values of $(^{147}Sm/^{144}Nd)_{CHUR}$ = 0.1967 and $(^{143}Nd/^{144}Nd)_{CHUR}$ = 0.512638; Jacobsen and Wasserburg, 1984) and ε_{Nd} (2.1 Ga) of the rocks, the $f^{source}_{Sm/Nd}$ values range from -0.09 to 0.12. The $(^{147}Sm/^{144}Nd)_{source}$ of the rocks has then been computed using the relationship below:

$$f_{Sm/Nd} = \frac{(^{147}Sm/^{144}Nd)_{source} - (^{147}Sm/^{144}Nd)_{CHUR}}{(^{147}Sm/^{144}Nd)_{CHUR}}$$

The $(^{147}Sm/^{144}Nd)_{source}$ obtained for the Type I basalts/basaltic andesites ranges from 0.209 to 0.220, which is about ~6–12% higher than the average chondrite. For the Type II basalts/basaltic andesites, the $(^{147}Sm/^{144}Nd)_{source}$ value ranges from 0.193 to 0.201, which is about ~2% lower to ~2% higher than the average chondrite. The $(^{147}Sm/^{144}Nd)_{source}$ values of the andesitic and dacitic rocks are from 0.190 to 0.197, representing ~4% lower to 0.4% higher than the average chondrite. The fractionation factor (á) was computed for the rocks using the expression of DePaolo (1988):

$$\alpha_{Sm/Nd} = \frac{1 + f^{rock}_{Sm/Nd}}{1 + f^{source}_{Sm/Nd}}$$

The á values obtained for the various rocks are presented in Table 2. The $á_{Sm/Nd}$ of the Type I basalts/basaltic andesites, Type II basalts/basaltic andesites, and the andesites and dacites indicate about 6–9%, 15–31% and 26–35% depletion in the Sm/Nd ratios of the precursors of these rocks at source, respectively.

The Sm–Nd isotopic modeling suggests that the volcanic were derived from heterogeneous mantle sources.

2. Tectonic setting

The geotectonic environment in which the rocks were formed or emplaced was investigated in the plot of $f_{Sm/Nd}$ versus ε_{Nd} (2.1 Ga). The Type I basalts/basaltic andesites plot in the field of MORB but close to the arc crust. Indeed, their ε_{Nd} (2.1 Ga) values overlap the fields of Paleoproterozoic arc crust and MORB (Fig.

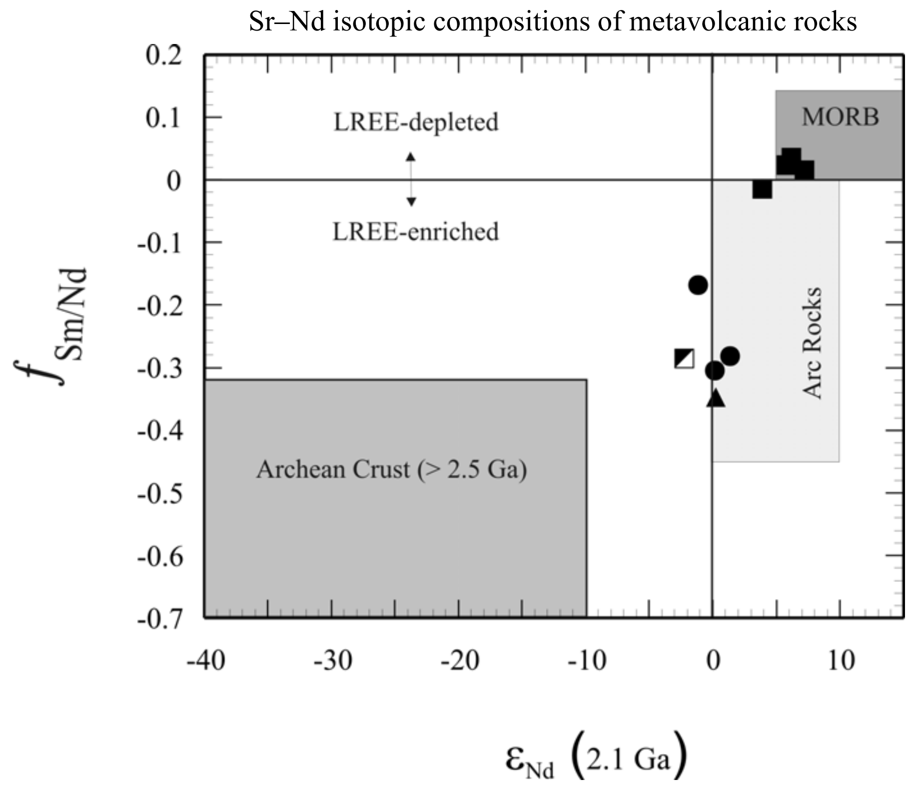

Fig. 6. Plot of ε_{Nd} (2.1 Ga) versus fractionation parameter ($f_{Sm/Nd}$). Fields of old (Archean) continental crust, MORB and the arc rocks are from Roddaz et al. (2007) and the references therein. Symbols as in Fig. 4.

6; Roddaz et al., 2000 and the references therein). The Type II basalts/basaltic andesites, andesites and dacites are either plotted in and close to the field of the Proterozoic arc crust.

The geotectonic setting of the Paleoproterozoic rocks of Ghana is contentious. Leube et al. (1990) have proposed an intracontinental rift model for the Paleoproterozoic rocks, whereas arc or subduction related models have been suggested by others (Sylvester and Attoh, 1992; Pohl and Carlson, 1993; Loh and Hirdes, 1999; Asiedu et al., 2004; Dampare et al., 2005; Attoh et al., 2006; Dampare et al., 2008, in review). Feybesse et al. (2006) have invoked a combination of continental margin, juvenile magmatism and convergence and collision between an old continent and a juvenile crust for the tectonic environments in which the Birimian greenstone belts were generated. For Harcouet et al. (2007), the basement in the Ashanti area is likely to be of continental type rather than oceanic, as the results of their numerical modeling using thermal parameters such as thermal conductivity values and heat-production rates are more consistent with the continental basement scenario. Sylvester and Attoh (1992) have indicated that the trace element chemistry of Birimian volcanic belts in Ghana is comparable to that of Archean greenstone belts, and that intermediate calc-alkaline units show the high Ba/La and Ta depletion which is characteristic of similar rocks in modern subduction environments and distinct from those formed in intraplate settings. The authors therefore suggest that Birimian volcanic rocks probably originated as immature island arcs built on oceanic crust. Dampare et al. (2008) have, on the basis of geochemistry, inferred an intra-oceanic island arc–fore-arc–back-arc setting for the Paleoproterozoic metavolcanic rocks from the southern Ashanti volcanic belt. The formation of most of the Paleoproterozoic volcanic belts in the oceanic context has also been underlined by other workers (e.g., Abouchami et al., 1990; Liégois et al. 1991; Boher et al., 1992; Davis et al., 1994; Vidal and Alric, 1994; Ama Salah et al., 1996; Béziat et al. 2000; Egal et al., 2002). Available geochronological and isotopic data (e.g., Abouchami et al. 1990; Liegeois et al. 1991; Boher et al. 1992; Taylor et al. 1992; Hirdes et al. 1996; Ama Salah et al. 1996; Doumbia et al. 1998; Gasquet et al. 2003) indicate a juvenile source for the Paleoproterozoic volcanic and granitic rocks, with the exceptions of plutons from the Winneba area in the southeast of

Ghana (Taylor et al., 1992) and the vicinity of the Man nucleus (e.g., Kouamelan et al., 1997), where there is evidence of a stronger influence of recycled Archean basement.

Abouchami et al. (1990) and Boher et al. (1992) have proposed a mantle plume model to explain crustal growth event in the West African Craton. According to the authors, extensive plateaus were formed by a mantle plume event, followed by formation of island arcs on the top of the oceanic plateaus which then collided with the Man Archean Craton. For the crustal growth event in Ghana, Taylor et al. (1992) have indicated that the Birimian represents a major early Proterozoic magmatic crust-forming event around 2.3–2.0 Ga by differentiation from a slightly depleted mantle source. The authors suggest that the Birimian of Ghana forms part of the major Proterozoic (Eburnean) episode of juvenile crustal accretion which is recognized in West African Craton and dated at 2.2–2.1 Ga by Abouchami et al. (1990). Davis et al. (1994), also following the accretionary model, have proposed that the Birimian sedimentary basins resulted from accretion of arcs and oceanic plateaus now represented by allochthonous Birimian volcanic belts. For Feybesse et al. (2006) the Birimian volcanic belts correspond to a major period of accretion of juvenile basic volcanic–plutonic rocks to an Archean continental domain around 2.25–2.17 Ga, whereas the Birimian sedimentary activity occurred at 2.15–2.10 Ga (Feybesse et al., 2006).

Our isotopic data favors a supra-subduction magmatism in the study area, consistent with the previously published geochemical data on the metavolcanics (Dampare et al., 2008) but inconsistent with the mantle plume model proposed for Paleoproterozoic rocks in Ghana (e.g., Leube et al., 1990). This is consistent with the island arc complex model that the Paleoproterozoic terranes of West Africa evolved through subduction-accretion processes.

IX. Conclusion

The metavolcanic rocks analyzed in this work are predominantly basalts/basaltic andesites and andesites with minor dacites. Two main basalts/basaltic andesites (B/A) have been identified, and are subsequently referred to as Type I B/A and Type II B/A. The tholeiitic Type I B/A are underlain by the calc-alkaline Type II B/A, andesites and dacites.

The analyzed metavolcanic rocks generally have low initial $^{87}Sr/^{86}Sr$ ratios values comparable to those previously reported on the Birimian basaltic and granitic rocks of West Africa. The isotope signatures of the Type I B/A suggest that their parent magma was derived from a depleted mantle source. They show high positive initial ε_{Nd} (T), i.e., +3.89 to +7.21. The wide range of epsilon values suggests some contribution from endogenic sources. The Type I B/A display Nd model ages (T_{DM2}) of 1.83–2.09 Ga similar to their formation ages, suggesting that they were juvenile at their time of formation. The andesites and Type II basalts/basaltic andesites show a range of ε_{Nd} (2.1 Ga) values from –1.15 to +1.35, which are less than those of Type I B/A. Their Nd model ages (T_{DM2} = 2.31–2.58 Ga) are also higher than the formation age (2.1 Ga), which suggest that they may have received some inputs from crustal sources. The andesites and andesites and dacites show a range of initial ε_{Nd} (T) values from –2.24 to +0.24, indicative of much higher crustal inputs or partly enriched to enriched mantle products. The initial ε_{Nd} value of the rock coupled with geochemical features could point to subduction-related lithospheric mantle sources. Their pre-Birimian Nd model age of 2.41–2.64 Ga are possibly indicative of a greater subduction input of an older crustal material. The trace element data, of the metavolcanic rocks argue against any major crustal contamination (Dampare et al., 2008). Therefore, the isotopic heterogeneities demonstrated by the rocks may probably be due to variable endogenic contamination (i.e., contamination within the mantle rather than the crust, such as inputs from subducted sediments).

The radiogenic isotopic signatures are consistent with the trace element data in suggesting a supra-subduction zone setting for the metavolcanic rocks from the study area. The Type I basalts/basaltic andesites were generated from a depleted mantle in a back-arc basin, whereas the arc-related Type I B/A, andesites and dacites were derived from heterogeneous, metasomatized lithospheric mantle sources. The isotopic signatures of the metavolcanic rocks support supra-subduction magmatism in the southern Ashanti greenstone belt of Ghana.

Acknowledgements

The field work was partly sponsored by the National Nuclear Research Institute (NNRI) of Ghana Atomic Energy Commission. The first author gratefully acknowledges the financial support received from the Japanese Government (Monbukagakusho: MEXT).

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OKAYAMA University
Earth Science Reports
Vol. 16, No. 1, 29-42, (2009)

Provenance of Early Cretaceous Hayama Formation, Okayama Prefecture, Inner Zone of Southwest Japan: constraints from modal mineralogy and mineral chemistry of derived detrital grains

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Petrographic and phase chemistry studies of detrital grains were carried out on sandstones from the Lower Cretaceous Hayama Formation, Inner Zone of Southwest Japan, to determine their provenance and the tectonic setting during the early Cretaceous. The results of the modal mineralogy suggest that the Hayama Formation has magmatic arc provenance and that deposition of the sediments took place in the back-arc areas with detritus mostly derived from the magmatic arc and rifted continental margins. The chemical compositions of chromian spinel, chlorite and sphene indicate that significant proportions of the detrital grains were derived from mafic and/or ultramafic sources. The source areas are the mafic and ophiolitic rocks in the Sangun-Renge and Akiyoshi terranes and the felsic volcanic rocks probably from either the Akiyoshi terrane or a source not presently exposed in southwest Japan. However, minor amounts of the detritus were derived from the basement rocks; i.e., carbonates and siliciclastic rocks of the Akiyoshi terrane and the metamorphosed mafic rocks of the Chizu terrane.

Keywords: modal analysis; sandstone; mineral chemistry; provenance; Hayama Formation.

I. Introduction

It is now well established that the composition of sedimentary rocks retains a record of geologic history (e.g., Dickinson and Suczek, 1979). The compositions of sedimentary rocks are commonly used as constrains on potential source areas (e.g., McLennan et al., 1995), to reconstruct tectonic settings (e.g., Dickinson et al., 1983), to understand crustal evolution (e.g., Taylor and McLennan, 1986; Condie, 1993) and to reveal paleoweathering and possible paleoclimatic conditions (e.g., Condie et al., 2000). These investigations are carried out using a combination of sandstone petrography and various geochemical and isotopic techniques.

This paper reports mineralogical data for the sandstones of the Lower Cretaceous Hayama Formation (Suzuki et al., 2001) that is distributed sporadically in Okayama and Hiroshima Prefecture, the Inner Zone of Southwest Japan (Fig. 1a). We discuss the provenance and tectonic setting of these sandstones, and attempt to reconstruct the paleogeography of the study area during the early Cretaceous. In this study we have dealt with two provenance variables, i.e., modal analysis of framework grains and phase chemistry of detrital mineral species; Asiedu et al. (2000a) have already dealt with the bulk-rock geochemical aspect of the provenance.

II. Geological overview

The Hayama Formation (about 400m thick) is composed mostly of limestone breccia, conglomerate, sandstone, and red shale, with minor thin alternating beds of sandstone and mudstone. The formation can be stratigraphically divided into the Upper and Lower members (Asiedu and Suzuki, 1995). The Lower Member (about 300m thick) is dominated by conglomerate and the Upper Member (about 100m thick) by mudstone. The Upper Member includes some

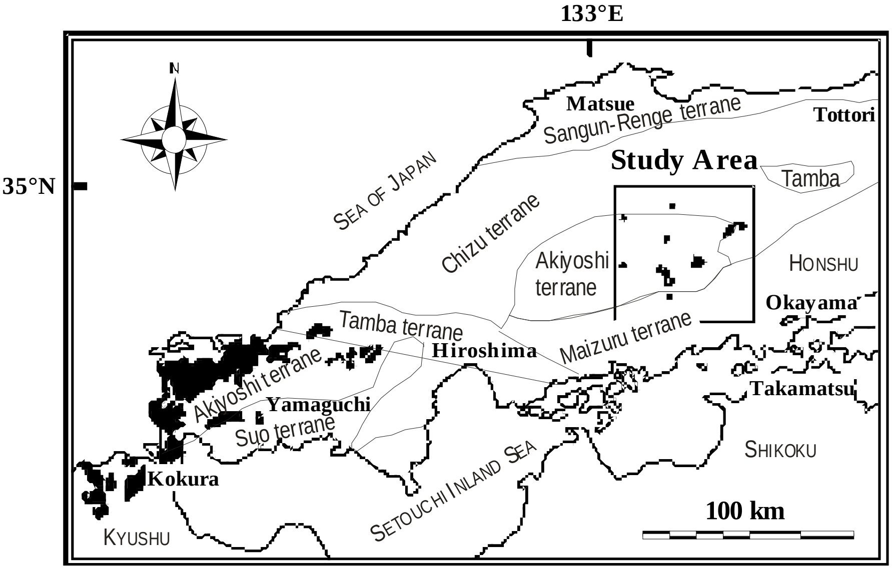

Fig. 1a. Map of inner side of Southwest Japan showing the location of the study area (Pre-Cretaceous terranes after Nishimura, 1990).

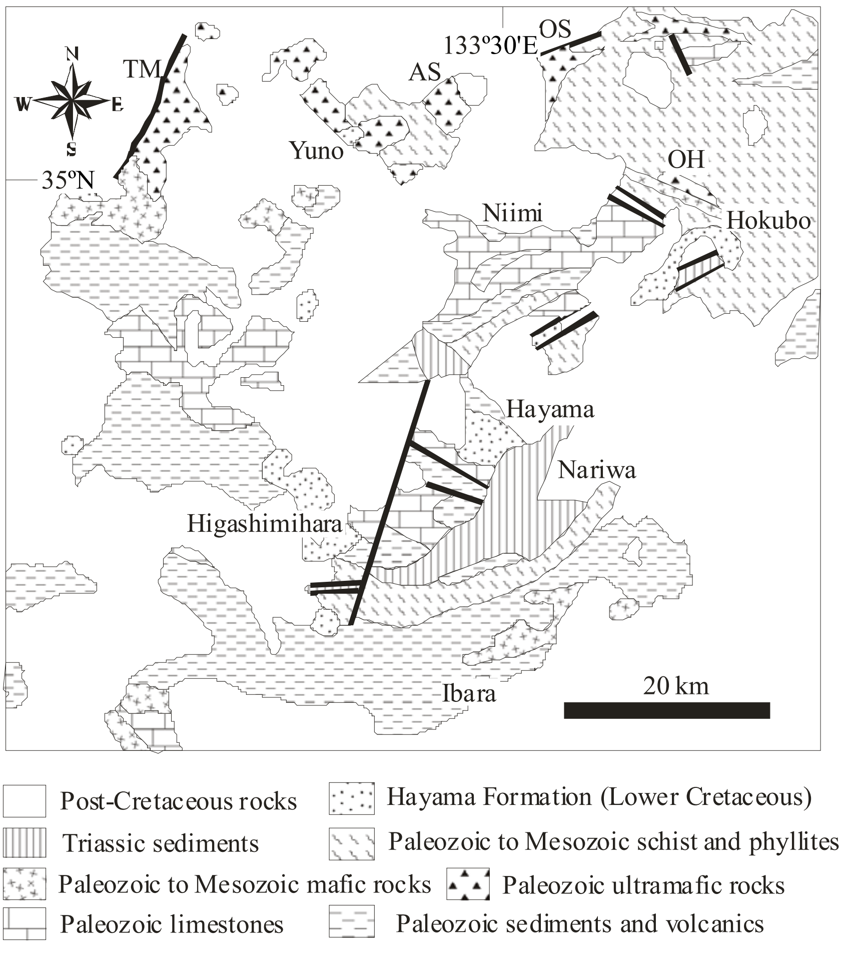

Fig. 2. Ternary diagram showing the compositions of the detrital feldspar grains from the sandstones of the Hayama Formation.

interbeds of felsic tuff. Fission-track dating on samples of the felsic tuff yielded an Albian age (Suzuki et al., 2001). The Formation correlates with part of Wakino Subgroup and Sasayama Group of southwest Japan, and the Kyeongsang Subgroup of southeastern Korea, on the basis of non-marine fauna and non-volcanic lithology. In the Hayama area, the formation is characterized by fining-upward successions with conglomerate at the base and mudstone at the top. Sedimentary facies analysis indicates a fluvial depositional environment and that the detritus was mostly derived from the north (Asiedu and Suzuki, 1995). Asiedu (1998) reconstructed the geometry of the sedimentary basin as a narrow valley.

The basement rocks of the Hayama Formation consist mainly of Paleozoic rocks with minor amounts of Triassic and Jurassic sedimentary rocks (Fig. 1b). They all belong to the Chizu and Akiyoshi terranes. The Paleozoic rocks are composed of limestone, sandstone, mudstone, chert, schist, and felsic and mafic tuffs. Pre-Cretaceous igneous rocks, which include ultramafic, gabbroic, doleritic and granitic rocks, intrude the Paleozoic rocks. The Triassic and Jurassic strata are non-marine to shallow marine in origin and are composed of sandstone and mudstone.

III. Laboratory methods

About 30 sandstone samples spanning the entire stratigraphic section were collected from the Hokubo, Hayama, and Higashimihara areas for a provenance study (Fig. 1). It was not difficult to sample along the entire stratigraphic section because the formation has undergone very little post-depositional tectonic deformation and bedding is near parallel, making statigraphic height proportional to contour levels. Fresh rock exposures are very abundant and readily accessible for sampling.

Seventeen medium-grained sandstone samples were selected for thin- and polished-section study. For each sample, two specimens comprising one thin-section and one polished-section, were prepared for modal analyses of framework grains and mineral chemistry of detrital grains, respectively. Modal analysis on the selected thin-sections was carried out using the Gazzi-Dickinson point-counting technique (Ingersoll et al., 1984). This involves counting grains greater than than 0.0625 mm as individual phases even where they form part of a lithic fragment. The purpose of this is to eliminate apparent variations in the composition of samples resulting only from differences in grain size. Between 400 and 600 points were counted per thin section.

The polished-sections were studied for mineral chemistry by microprobe technique. The phase chemistry of grains of selected minerals (i.e., chromian spinel, epidote, sphene, chlorite and feldspar) was examined with a three-channel JOEL JXA-733 microprobe analyzer at the Department of Earth Sciences, Okayama University. The operating conditions were as follows: acceleration voltage, 15 kV; probe current, 20 nA; counting time, 10 seconds per element; and beam diameter, 10 μm.

IV. Mineralogical constituents

The results of the point-count are shown in Table 1. No significant compositional variations with stratigraphy were observed though the samples nearer to the base of the Formation are slightly richer in chlorite and heavy minerals. The sandstones are texturally and mineralogically immature, and have undergone very little or no metamorphism. The sandstone compositions are characterized by high proportion matrix (between 15 and 30 vol.%; Table 1). According to the sandstone classification of Folk (1974) they are lithic graywackes.

1. Quartz

The proportion of the quartz (including microcrystalline chert) ranges from 5 to 15% of the detritus. Monocrystalline quartz is slightly more abundant than the polycrystalline variety (Table 1). Most of the monocrystalline quartz grains are of volcanic origin as they typically show euhedral shapes, non-undulatory extinction, embayments, and inclusion-free clear transparency. Such volcanic quartz is present in all the analyzed samples. Monocrystalline quartz crystals showing tectonic fabric mostly contains mineral inclusions of white mica. Polycrystalline quartz grains are composed mainly of non-oriented crystallites, commonly three or more crystals per grain, with straight to undulose extinction and straight grain boundaries.

Table 1. Point-count data for the sandstones of the Hayama Formation

Sample No	Qm	Qp	K	P	Lv	Lm	Ls	Chl	HM	C	Matrix
Hokubo area											
UM-H-06	6.3	4.2	3.4	18.4	24.3	1.2	7.6	11.4	1.9	1.9	19.2
UM-H-05	5.5	5.5	0.0	13.8	20.3	6.8	5.9	10.2	1.8	3.4	26.8
LM-H-01	5.9	4.1	1.0	14.8	24.6	4.1	6.5	9.8	3.1	3.2	22.9
LM-H-02	6.1	4.1	1.3	14.5	25.8	3.6	6.2	9.4	3.3	3.0	22.8
LM-H-04	9.8	4.6	1.4	15.0	22.5	2.8	5.2	12.3	1.9	3.4	21.5
LM-H-03	7.4	5.3	2.1	12.0	22.9	4.4	5.2	7.3	2.2	1.9	29.3
Higashimihara area											
UM-I-03	5.8	5.5	0.0	14.2	24.9	1.0	9.0	9.7	3.4	3.3	23.3
LM-I-01	4.7	3.9	0.4	13.1	26.8	1.7	6.8	15.7	3.7	4.3	18.9
LM-I-02	2.4	2.8	0.0	11.7	31.0	7.0	6.5	9.6	3.0	2.8	23.0
Hayama area											
UM-N-08	5.1	3.9	0.2	19.2	20.1	4.8	2.1	9.2	2.9	3.5	28.9
UM-N-07	8.7	2.4	0.9	5.2	39.9	0.2	9.0	1.0	0.5	4.4	27.8
UM-N-06	6.9	3.7	3.0	21.5	21.5	5.3	5.9	4.0	2.3	0.7	25.1
UM-N-05	8.2	5.9	3.1	15.8	27.1	3.9	4.2	6.2	2.5	2.6	20.6
LM-N-01	6.3	3.7	2.9	14.4	30.7	0.2	5.4	8.4	7.8	0.6	19.5
LM-N-02	4.8	3.6	2.4	17.4	17.7	3.8	6.7	14.8	4.7	2.6	21.5
LM-N-03	8.6	4.3	1.1	13.8	23.9	4.3	6.3	14.3	3.4	4.6	15.4
LM-N-04	7.9	5.2	1.2	20.8	25.1	1.9	7.1	5.0	2.1	3.2	20.7

Abbreviations: Qm = monocrystalline quartz; Qp = polycrstalline quartz including chert; P = plagioclase feldspar; F = potassium feldspar; Lm = metamorphic lithic grains; Lv = volcanic lithic grains; Ls = sedimentary lithic grains; Chl = chlorite; C = calcite; HM = heavy mineral; M = matrix content; LM-, Lower Member; UM-, Upper Member

2. Feldspar

Euhedral to subhedral feldspar grains constitute between 6 and 24% of the detritus. Plagioclase is by far the most abundant feldspar variety (Table 1). Both twinned and untwinned plagioclase varieties occur but the latter is more abundant. Plagioclase may be fresh and unaltered but, more commonly, they are replaced by carbonate or altered to clay minerals. Potassium feldspar, which includes sanidine and orthoclase, was distinguished from quartz by presence of cleavage, cloudy alteration and lower refractive indices. Microcline and microperthite were observed in few samples. Microprobe analysis carried out on fresh feldspar crystals with little or no replacement textures shows the dominance of albite and sanidine (Table 2; Fig. 2). The use of feldspar grains in provenance studies, however, is greatly hampered by the process of albitization during sediment diagenesis. To minimize the effect of albitization, only detrital feldspars with no replacement textures were analyzed. The detrital feldspars show bimodal distribution, even in a single thin-section. Dutta and Wheat (1993) have suggested that such compositional bimodality may indicate that the feldspar composition is primary.

3. Rock fragments

Lithic volcanic clasts are the most predominant component in all the analyzed samples (17 to 30%). Two broad categories were identified: felsic volcanic and mafic volcanic clasts. The felsic volcanic clasts are more abundant and are characterized by microcrystalline mosaic of individual quartz crystals with relict feldspar. Vitric varieties also occur and have a groundmass consisting mainly of altered and devitrified glass. Relict shards and flow structures were observed in some samples. Mafic volcanic fragments

Table 2. Representative compositions of feldspar (single grains)

Sample No	LM-N-01		UM-N-08	LM-H-02	
Analysis No	1	13	5	2	7
SiO_2	66.84	69.56	64.34	69.76	68.27
Al_2O_3	20.61	18.68	18.34	18.49	21.37
CaO	2.49	0.54	0.19	0.18	0.72
Na_2O	10.20	11.09	0.81	11.45	9.48
K_2O	0.09	0.07	15.45	0.05	0.61
FeO	0.14	0.06	nd	0.16	nd
Total	100.37	100.00	99.13	100.09	100.45
Formula on the basis of 32 oxygens					
Si	11.698	12.126	11.974	12.154	11.830
Al	4.251	3.838	4.023	3.797	4.364
Ca	0.467	0.101	0.038	0.034	0.134
Na	3.461	3.748	0.292	3.868	3.185
K	0.020	0.016	3.668	0.011	0.135
Fe	0.020	0.009	nd	0.023	nd
Total	19.917	19.838	19.995	19.887	19.648
Molecular Ratio					
Ab	87.66	96.97	7.30	98.85	92.21
An	11.83	2.61	0.95	0.87	3.88
Or	0.51	0.41	91.75	0.28	3.91

nd, not determined

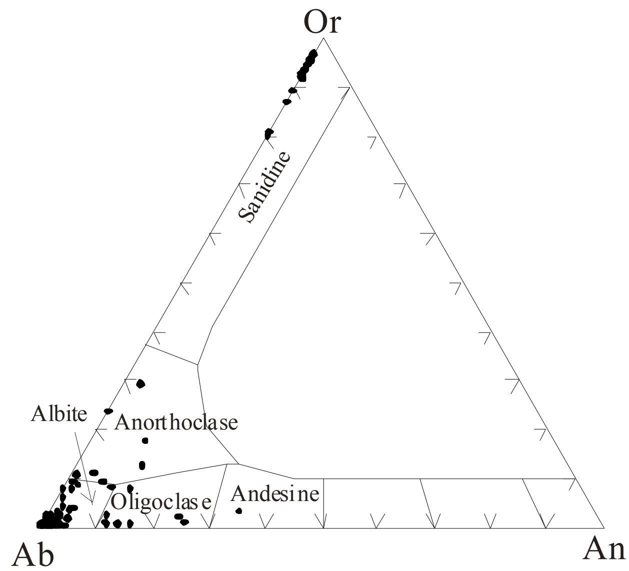

Fig. 2. Ternary diagram showing the compositions of the detrital feldspar grains from the sandstones of the Hayama

mostly show a microlitic texture with laths of plagioclase in an altered alphanitic groundmass. Irregular opaque fragments have been interpreted as volcanic mafic fragments due to occasional appearance of plagioclase and mafic mineral inclusions.

Sedimentary rock fragments include sandstone, mudstone, chert and calcareous fragments. Chert grains include both microcrystalline and radiolarian varieties; microcrystalline chert is, however, included in the polycrystalline quartz category for the modal analysis (Ingersoll et al., 1984). Microcrystalline chert grains were distinguished from microcrystalline felsic volcanic rock fragments by a lack of marked internal relief between individual crystals, lack of feldspar microphenocrysts and sometimes by presence of criss-crossing veinlets. Detrital calcite grains, which constitute up to 3% of the detritus, often contain fragments of fusulinids and other Paleozoic

Table 3. Representative compositions of chlorite aggregate

Sample No	LM-N-02		LM-H-02		
Analysis No	4	7	1	2	8
MgO	17.58	25.26	20.63	32.05	25.60
FeO(tot)	17.90	12.68	14.13	5.64	11.66
SiO_2	28.06	30.74	30.25	31.20	30.41
Al_2O_3	22.07	13.99	18.79	16.13	18.84
Cr_2O_3	0.22	2.03	0.75	0.54	0.09
Ni	0.20	nd	0.15	0.17	0.14
Total	86.03	84.70	84.70	85.73	86.74
(H_2O)	13.97	15.31	15.70	14.27	13.27
Formula on the basis of 28 oxygens					
Mg	5.367	7.704	6.272	9.278	7.512
Fe	3.065	2.169	2.410	0.916	1.919
Si	5.746	6.289	6.169	6.059	5.985
Al	5.326	3.373	4.516	3.692	4.370
Cr	0.036	0.328	0.121	0.083	0.014
Ni	0.033	nd	0.025	0.027	0.022

nd, not determined

foraminiferas similar to those of the Paleozoic limestones distributed in the study area.

Metamorphic rock fragments are generally poorly represented in the sandstones. They are composed of low-grade metamorphic fragments, most probably derived from schists.

4. Phyllosilicates (including chlorites)

White mica is the most dominant phyllosilicate mineral in all the analyzed sandstones and occurs mainly as short plates. It occurs as free grains, as inclusions in quartz and feldspars, and as constituents in rock fragments. Biotite is generally rare.

Chlorite is a common mineral in the sandstones and accounts for up to 15% of the detrital constituents. In thin sections it mainly occurs as single anhedral large flakes with extremely irregular outlines and shows brown or navy blue interference colors. Some of the grains that show navy blue interference colors contain inclusions of opaque minerals and chromian spinel. The occurrence of chromian spinel as inclusions in the navy-blue chlorites together with their chemistry (MgO 16 to 32 %; Cr up to 2030 ppm; Ni up to 2150 ppm; Table 3) suggests a mafic/ultramafic source for these chlorites (Wrafter and Graham, 1989). The navy-blue chlorites can be classified as clinochlore and/or pycnochlorite, and the brown as brunsvigite and/or diabantite (Fig. 3). Although most of the chlorites appear detrital in origin, they were not included in the volcanic lithic grains component of the point counts.

5. Heavy minerals

The sandstones are rich in heavy minerals; their proportion ranges from 0.5 to 8 percent of the detrital content. The heavy mineral assemblage includes, in order of decreasing abundance, epidote, opaque minerals (mostly magnetite and subordinately ilmenite and pyrite), chlorite, chromian spinel, sphene, anatase, zircon and garnet. The zircon grains are mostly either rounded or angular in shape; euhedral types are generally rare. The garnet grains are angular, colorless and truly isotropic. Mineral chemistry of 3 detrital grains shows that they are pyrope-almandine in composition.

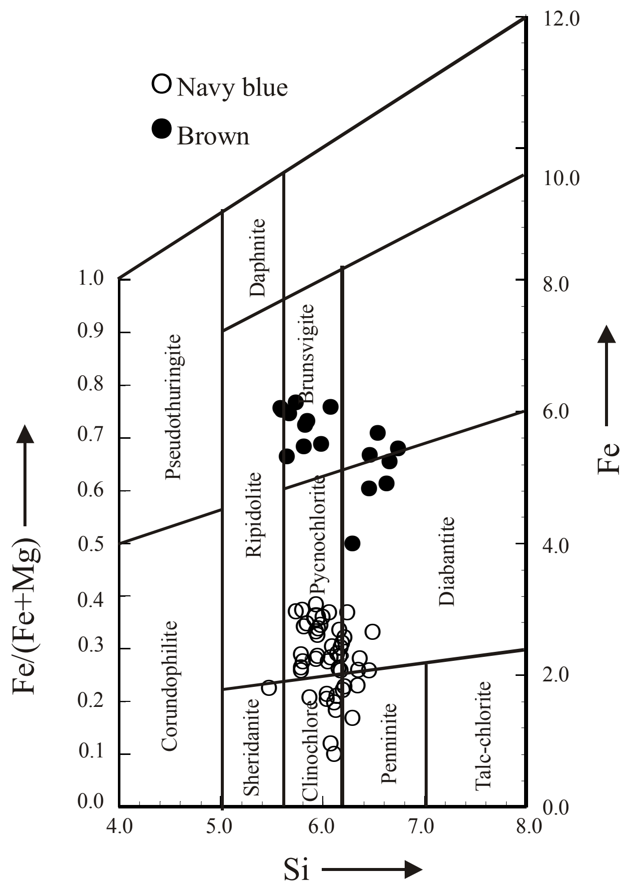

Fig. 3. Chemical compositions of chlorite aggregates from the Hayama sandstones (fields after Hay 1954).

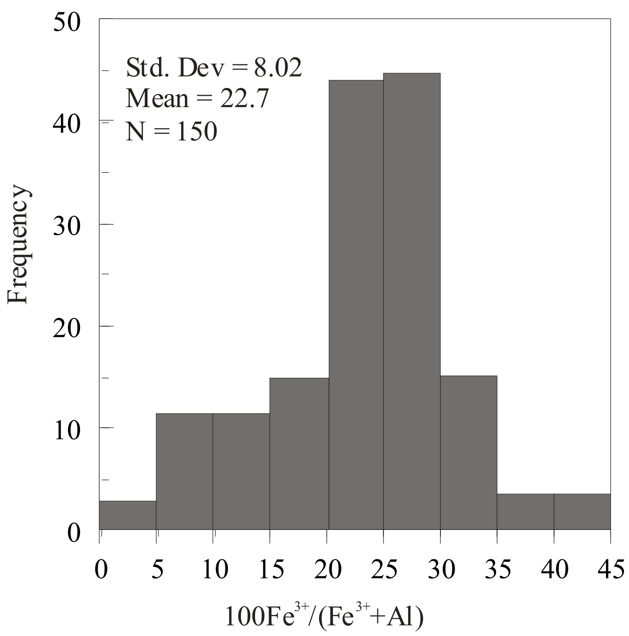

Fig. 4. Frequency of $Fe^{3+}/(Fe^{3+}+Al)$ ratios of detrital epidote from the Hayama sandstones.

a. Epidote

Epidote accounts for over 65% of the heavy mineral suites in some samples. It is pale green or yellowish green in color and is mostly irregular, angular or equant in shape. Few colorless grains were encountered in the petrographic study. The Fe/(Fe+Al) ratios are useful in characterizing epidotes. The epidote grains from the analyzed samples show a wide scattering in their Fe/(Fe+Al) ratios with a mean of about 0.22 (Table 4; Fig. 4). The epidote compositions are comparable to those reported from metagabbros and epidote-amphibolites in the Oeyama ophiolite (Kurokawa, 1985).

b. Chromian spinel

The chromian spinel grains are dark reddish brown to yellowish brown in color and generally show sharp angular shapes with irregular outlines. They occur in almost all the analyzed samples but are generally richer in the samples from the Lower Member of the formation. The chemistry of the detrital chromian spinel suggests derivation from ultramafic members of an ophiolite suite and is discussed in detail by Asiedu et al. (1997). The compositional range of the detrital spinel is similar to those from the ultramafic bodies distributed in the Akiyoshi terrane (Table 5; Fig. 5).

c. Sphene

The detrital sphene grains are mostly pale green in color, euhedral to irregular in shape and mostly free of inclusions. The chemical compositions are very homogeneous and suggest their derivation from igneous rocks (Table 6; Fig. 6). Sphenes from acidic and intermediate igneous rocks contain appreciable amounts of Fe, Al, rare earth elements, and Y whereas those from basic and/or ultrabasic rocks have low or negligible Fe, Al, rare earth elements and Y concentrations. The sphene analyses of the Hayama sandstones show compositions close to the ideal cation proportion of $CaTiSiO_5$, suggesting their derivation from basic and/or ultrabasic rocks (Asiedu et al., 2000b).

Table 4. Representative compositions of detrital epidote

Sample No	UM-H-05	UM-N-08	LM-N-02	LM-N-01	LM-I-01
Analysis No	2	4	1	3	3
SiO_2	38.07	38.07	38.93	38.62	37.83
TiO_2	0.11	0.01	0.24	0.00	0.02
Al_2O_3	25.68	25.63	27.66	28.26	22.01
Cr_2O_3	0.02	0.03	0.02	nd	nd
Fe_2O_3	11.20	11.59	9.57	6.50	13.42
MnO	0.20	0.09	0.11	nd	0.02
MgO	0.02	0.02	1.40	nd	nd
CaO	23.85	23.84	21.60	24.20	23.25
NiO	0.01	0.00	0.01	nd	nd
Total	99.16	99.28	99.54	97.57	96.61
Formula on the basis of 12.5 oxygens					
Si	3.091	3.089	3.100	3.045	3.136
Ti	0.007	0.001	0.014	0.000	0.001
Al	2.457	2.451	2.596	2.626	2.150
Cr	0.001	0.002	0.001	nd	nd
Fe^{3+}	0.684	0.708	0.573	0.429	0.930
Mn	0.014	0.006	0.007	nd	0.006
Mg	0.002	0.002	0.166	nd	nd
Ca	2.074	2.072	1.843	2.044	2.065
Ni	0.001	0.000	0.001	nd	nd

nd, not determined

Table 5. Representative compositions of chromian spinel

Sample No	LM-N-01	UM-N-08	LM-H-01	UM-H-05	LM-H-02
Analysis No	3	5	1	1	2
TiO_2	0.00	0.04	0.04	0.05	0.00
Al_2O_3	24.72	30.27	21.46	24.35	31.72
Cr_2O_3	43.19	38.59	47.73	45.75	36.82
Fe_2O_3	1.07	0.82	0.82	0.49	0.69
FeO	17.36	14.78	17.17	15.34	15.78
MnO	0.28	0.25	0.31	0.26	0.22
MgO	12.60	15.11	12.35	13.76	14.32
Total	99.22	99.86	99.87	100.00	99.55
Formula on the basis of 32 oxygens					
Ti	0.000	0.007	0.007	0.009	0.000
Al	7.190	8.434	6.296	6.984	8.845
Cr	8.427	7.213	9.394	8.803	6.888
Fe^{3+}	0.198	0.147	0.153	0.090	0.123
Fe^{2+}	3.583	2.922	3.573	3.122	3.122
Mn	0.059	0.050	0.065	0.054	0.044
Mg	4.635	5.324	4.582	4.991	5.050
Cr/(Cr+Al)	0.54	0.46	0.60	0.56	0.44
$Mg/(Mg+Fe^{2+})$	0.56	0.65	0.56	0.62	0.62

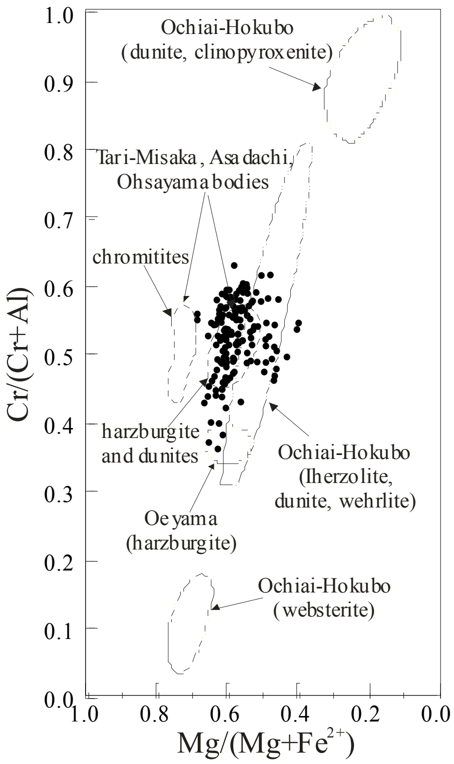

Fig. 5. Cr/(Cr+Al) versus Mg/(Mg+Fe^{2+}) plot of detrital spinel from the Hayama sandstones. Fields of the Tari-Misaka, Ashidachi and Ohsayama ultramafic bodies after Arai (1980) and Nozaka & Shibata (1994), those from

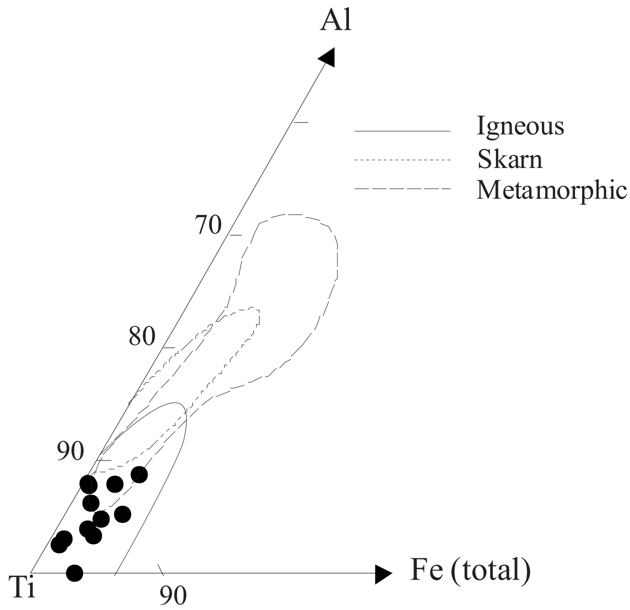

Fig. 6. Al – Ti – Fe (total) ternary plots of detrital sphene from the Hayama sandstones (fields after Asiedu et al. 2000).

Table 6. Representative compositions of detrital sphene

Sample No	LM-H-04		LM-I-01		LM-N-02
Analysis No	1	2	1	2	1
SiO_2	30.89	30.52	30.44	30.50	30.61
TiO_2	35.25	38.98	39.48	39.61	38.31
Al_2O_3	2.27	0.81	0.03	0.69	1.67
Fe_2O_3	0.19	0.01	0.03	0.00	0.02
FeO	1.28	0.39	1.22	0.36	0.57
MnO	0.01	0.13	0.00	0.02	0.00
CaO	27.66	27.56	27.53	27.72	27.52
Total	97.55	98.39	98.73	98.90	98.70
Formula on the basis of 20 oxygens					
Si	4.126	4.041	4.036	4.018	4.034
Ti	3.540	3.881	3.936	3.924	3.796
Al	0.357	0.126	0.005	0.107	0.259
Fe^{3+}	0.020	0.001	0.003	0.000	0.002
Fe^{2+}	0.142	0.043	0.136	0.040	0.063
Mn	0.001	0.015	0.000	0.002	0.000
Ca	3.958	3.909	3.910	3.913	3.885

6. Matrix

The mean matrix content of the analyzed sandstone samples from the Hayama Formation is 23%. The sandstone matrix is generally composed of fine-grained quartz (grain size <0.03 mm), chlorite, mica, opaques, altered lithic and feldspathic fragments, and material too fine to be identified. Although care was taken to avoid alteration due to weathering by collecting only the fresh samples, the high matrix contents suggest that diagenetic alteration of some grains to matrix may have occurred.

V. Discussion

1. Provenance

The petrographic and phase chemistry studies of the Hayama sandstones indicate a compositional mixture of at least four different detrital sources. A major part (estimated up to 50%) of the detritus is interpreted as the component derived from an ophiolitic source. Evidence for this interpretation is supported by the presence of detrital chromian spinel and mafic volcanic fragments and the chemistry of chlorite and sphene. This interpretation is in agreement with geochemical data that show concentrations of Mg, Cr, Ni, Co, Sc (Asiedu et al., 2000a). The lack of detrital pyroxene and amphibole and the abundance of Mg-rich chlorite and albite suggest that the ophiolitic source area may have experienced metamorphism that changed the primary rock constituents to epidote - chlorite - albite assemblages. The chemical compositions of the chromian spinel are comparable to those from ultramafic rocks within the Akiyoshi terrane (Fig. 5). The ultramafic and minor mafic rocks distributed in the Akiyoshi and Sangun-Renge terranes, or an ophiolite complex with a similar character, are therefore, the most probable source for this provenance component.

A significant proportion (estimated up to 40%) of the detritus is interpreted as the component derived from volcanic rocks of felsic composition. Evidence for this interpretation is supported by the abundance of felsic volcanic fragments, occurrence of quartz of volcanic origin and the chemistry of feldspar grains. Felsic volcanic rocks are sporadically distributed in the Akiyoshi terrane and are possible candidate for the supply of these detritus. However, the high proportion of felsic volcanic fragments in the analyzed sandstones, suggests that the source region must be a terrane with predominantly felsic volcanic composition. Therefore, if the Akiyoshi terrane is the source of this provenance component, then we can assume that large quantities of felsic volcanic rocks were distributed in the Akiyoshi terrane during the early Cretaceous. Alternatively, a silicic magmatic arc could have been the source of this provenance component. However, no obvious volcanic terrane of pre-late Cretaceous age is presently exposed in southwest Japan.

A third sedimentary source which contributed only a small amount of the total debris can be assumed for the mudstone fragments, carbonate fragments with associated foraminiferan fossils, red chert with radiolarian remains, and pyrite crystals. The limestone, chert, and mudstone-sandstone units of the Akiyoshi terrane are the probable source for this component.

A heterogeneous source characterizes the final estimated 10% of the detrital grains. This is evidenced by the presence of strained quartz with mineral inclusions, polycrystalline quartz, white mica, K-feldspar, metamorphic fragments, and basic volcanic fragments. The heterogeneous Chizu terrane is a potential supplier of this component.

The common association of the four provenance types in the investigated sandstones, together textural and mineralogical immaturity of the analyzed samples argues for a close proximity of the source areas. Dickinson and co-workers have related detrital sandstone compositions to major provenance types in various discrimination diagrams (Dickinson and Suczek, 1979; Dickinson et al., 1983). In several provenance studies (e.g., Dickinson et al., 1983), samples with greater than 25% matrix are disregarded in order to prevent mistaken provenance assignment due to diagenetic changes to the modal composition. Although 5 out of the 17 Hayama Formation samples fall within this matrix limit, in order to provide data for the entire stratigraphic level, we have included all the 17 analyzed samples in the discrimination plots. In the ternary diagram of Dickinson et al. (1983) the analyzed samples plot in the magmatic arc field (Fig. 7a, b). Taking into consideration the inferred close proximity of the source area to the depositional site, we envisage that a continental magmatic arc presumably supplied the detritus to sedimentary basin where the Hayama Formation formed. The most reasonable geotectonic setting that accommodates all these provenance sources is continental basins formed in the back-arc region. Ingersoll and Suczek (1979) have designed triangular diagrams that permit differentiation of the tectonic settings of depositional basins. On these diagrams the analyzed samples plot in and around the back-arc basin field that is typified by the mixture of detritus from a magmatic arc and rifted continental margin (Fig. 7c, d).

2. Paleogeographic reconstruction

Okada and Sakai (1993) have indicated that the Inner Zone of Southwest Japan is a typical active continental margin characterized by back-arc sedimentary basins controlled by strike-slip faults that are related to the Nagato tectonic zone. Early Cretaceous sedimentary basins in Southwest Japan are closely related to these strike-slip faults and can be classified as strike-slip basins with half-graben structure (Okada and Sakai, 1993; Sakai and Okada, 1997). The Hayama Formation, however, may represent a typical back-arc basin; the basin is not bounded or controlled by fault (Asiedu, 1998). In addition, the sandstone compositions of the Hayama Formation compares well with those from modern

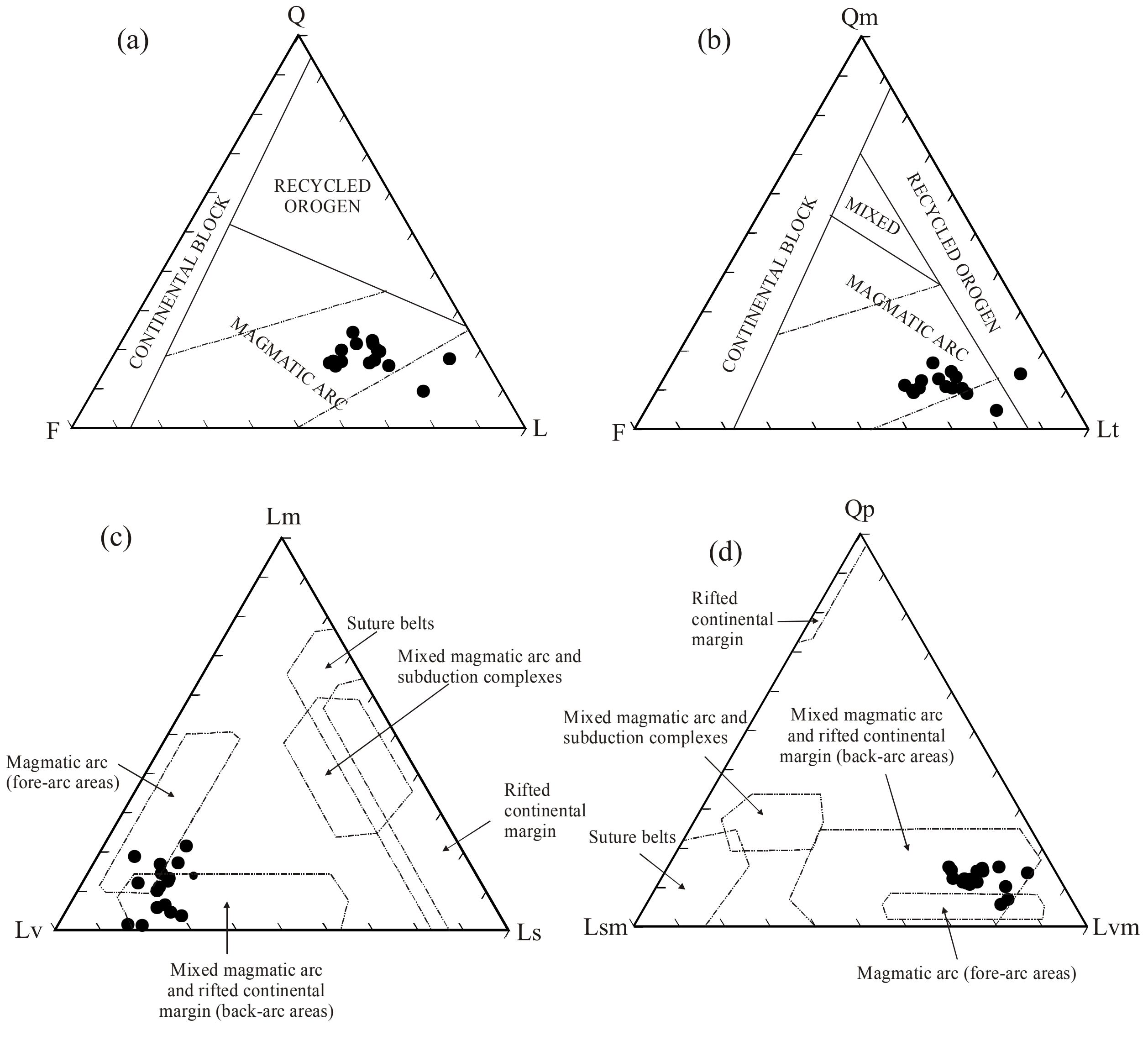

Fig. 7. Ternary diagrams for the analyzed samples: (a) Q–F–L and (b) Qm–F–Lt, after Dickinson et al. (1983); (c) Lm–Lv–Ls and (d) Qp–Lsm–Lvm, after Ingersoll and Suczek (1979). Abbreviations same as Table 1; Lsm, lithic metasedimentary; Lvm, lithic metavolcanic.

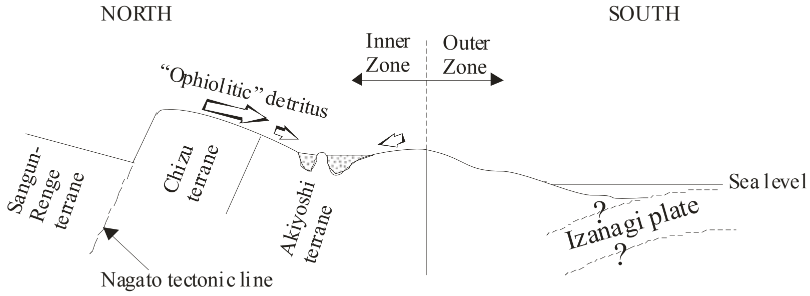

Fig. 8. Schematic illustration of provenance and tectonic environment at the time of deposition of the Hayama Formation (M.T.L = Median Tectinic Line). No scale implied.

back-arc basins and plots within the back-arc basin field of Ingersoll and Suczek (1979).

Okada and Sakai (1993) have suggested that the Lower Cretaceous sedimentary basins began to form at the time of oblique subduction of the Izanagi plate beneath the eastern margin of the Asian continent. Early Cretaceous magmatic rocks are not presently exposed in Southwest Japan but late Cretaceous magmatic rocks are widely exposed. This led Takahashi (1983) and Sakai and Okada (1997) to conclude that the nature of plate boundary during the early Cretaceous was of an oblique-slip type and that plate subduction and arc magmatism occurred in Southwest Japan in the late Cretaceous. It is, therefore, most likely that the volcanic rock fragments of felsic composition observed in the analyzed samples were supplied by felsic paleo-volcanic rocks rather than juvenile magmatic rocks.

We present a tectonic model for the study area as follows: The oblique subduction of the Izanagi plate resulted in a strike-slip movement along the Nagato tectonic zone located in the adjacent plate interior. The strike-slip motion caused a relative subsidence of the block in the interior side and an uplift of the block on the continental side. The subsiding block tilted to the south and sedimentary basins formed on the subsiding block show a half-graben structure (Okada and Sakai, 1993). The Hayama Formation is located to the south of the Nagato tectonic zone and therefore is located in the elevated block. The inferred southerly paleocurrent direction at the time of deposition of the Hayama Formation (Asiedu and Suzuki, 1995) suggests that the elevated block also tilted to the south, resulting in the obduction of ophiolitic rocks in the Sangun-Renge and Akiyoshi terranes for subsequent weathering, erosion and transport to the depositional site (Fig. 8). The heterogeneous basement rocks, and possibly a felsic magmatic arc located to the south of the basin, also supplied the rest of the detritus to the basin.

VI. Conclusions

The provenance of the Hayama Formation has been assessed using an integrated petrographical and phase chemistry approach of derived detrital grains. This approach has revealed that the sandstones of the Lower Cretaceous Hayama Formation is mineralogically immature and hetrolithic in nature and that the major sources are ultramafic rocks and felsic volcanic rocks. The provenance characteristics suggest that the Hayama Formation was deposited on the continental back-arc region of the magmatic arc that was silicic in composition. Major contributors of detritus to the depositional basin were the mafic and ultramafic rocks of the Sangun-Renge and Akiyoshi terranes and a felsic source whose location has not been positively identified.

Acknowledgements

This study was funded by the Japanese Ministry of Education, Science, Culture and Sports (to DKA).

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岡山大学地球科学研究報告
第 16 巻第 1 号*　2009 年　（年 1 回発行）
平成 21 年 12 月 25 日　発行

発行人　岡山大学地球科学研究報告編集委員会
塚本　修
e-mail: tsuka@cc.okayama-u.ac.jp

発行所　岡山大学理学部地球科学科
〒700-8530　岡山市北区津島中三丁目 1-1
TEL. 086 (251) 7891　FAX. 086 (251) 7895
http://www.desc.okayama-u.ac.jp/

* 年 1 回発行を原則とする．同一年度の追加発行は第 2 号以下の番号を使用.

printed in Japan

岡山大学
地球科学研究報告
2009年第16巻第1号

目次

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