[AMERICAN JOURNAL OF SCIENCE, VOL. 293, APRIL, 1993, P. 322–359]

POLYMETAMORPHIC EVOLUTION OF THE WESTERN BLUE RIDGE: EVIDENCE FROM $^{40}Ar/^{39}Ar$ WHOLE-ROCK SLATE/PHYLLITE AND MUSCOVITE AGES

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ABSTRACT. The importance of Ordovician tectonothermal activity in the western Blue Ridge of the southern Appalachians has been questioned by recent reports of Late Devonian-earliest Mississippian fossils within regionally metamorphosed rocks. In addition, metamorphism of fossiliferous Early Devonian rocks within the Talladega belt and suggested stratigraphic correlations with rocks of the Murphy belt suggest only post-Silurian metamorphism. The recent reports are contrary to most previous geochronology that suggests Ordovician metamorphism, as well as stratigraphic evidence indicating a Late Proterozoic age for most western Blue Ridge protoliths. To evaluate these contradictory results, eleven whole-rock samples (chlorite to garnet zones) and three muscovite concentrates (staurolite and kyanite zones) from the eastern Great Smoky Mountains of the western Blue Ridge were analyzed with $^{40}Ar/^{39}Ar$ techniques. Most chlorite-grade samples record plateau and intermediate temperature ages of 440 to 460 Ma. Illite crystallinity characteristics indicate that these samples attained metamorphic conditions sufficient for complete rejuvenation of whole-rock systems. Most biotite- and garnet-grade whole-rock samples yield plateau and intermediate temperature ages of 340 to 350 Ma. Muscovite samples record plateau ages of 360 to 380 Ma. It is unlikely that whole-rock samples collected several kilometers apart could have experienced contrasting cooling histories resulting in 100 Ma differences in apparent age. Therefore, the $^{40}Ar/^{39}Ar$ results most likely indicate a polymetamorphic history in which a 440 to 460 Ma (Middle to Late Ordovician) event was overprinted by a 360 to 380 Ma (Middle to Late Devonian) event. This interpretation is consistent with metamorphic textures observed in the western Blue Ridge.

INTRODUCTION

Multiple periods of metamorphism and deformation are well documented in the northern Appalachians by several lines of evidence. Textures indicative of polymetamorphism are common (Rosenfeld, 1968; Laird and Albee, 1981), and geochronologic evidence for Ordovician, Devonian, and Carboniferous events has been presented (Dallmeyer, 1982; Laird, Lanphere, and Albee, 1984; Hames and others, 1991). Furthermore, geochrologic results are compatible with paleontologic controls (Billings and Cleaves, 1934; Brookins, Berdan, and Stewart, 1973; Lyons and Darrah, 1978), and a clear sedimentary expression of these events is preserved in the Appalachian foreland (Colton, 1970; Hatcher and others, 1990). The record of deformation and metamorphism within the crystalline southern Appalachians, however, is not as

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well established. In large part this reflects erosion of most sedimentary cover sequences, rarity of documented fossils within metasedimentary rocks, and telescoping of metamorphic assemblages by extensive late Paleozoic thrust faulting.

Previous geochronologic studies in the southern Appalachians have suggested that most metamorphism and ductile deformation occurred in the Ordovician ("Taconic"). Devonian ("Acadian") and younger effects were believed to be recorded only in localized, retrogressive overprinting of earlier ductile fabrics. The importance of Ordovician metamorphism in this area, however, has recently been questioned because of reported fossil discoveries in the western Blue Ridge (Unrug and Unrug, 1990; Unrug, Unrug, and Palmes, 1991; Tull and others, in press) and the Talladega belt (Tull and others, 1988) that indicate protoliths of some metamorphosed clastic rocks were deposited during the early to middle Paleozoic. In addition, lithostratigraphic correlations between rocks exposed in the Murphy belt (western Blue Ridge) and fossiliferous Early Devonian strata in the adjacent Talladega belt have been suggested (Tull and Guthrie, 1985; Tull and Groszos, 1988). These interpretations imply that all deformation and metamorphism in the western Blue Ridge must be middle Paleozoic or younger. This is contrary to much of the previous geochronology reported for the southern Appalachians and a stratigraphic record of Ordovician tectonic instability in the Appalachian foreland.

In an attempt to evaluate these apparently contradictory results, a program of $^{40}Ar/^{39}Ar$ geochronology and structural analysis was undertaken in the Great Smoky Mountains of the western Blue Ridge. Samples from the lowest grade rocks exposed in the western Blue Ridge were analyzed to minimize potential effects of post-metamorphic cooling. These results are presented here and provide important constraints on the metamorphic evolution of the western Blue Ridge.

Paleozoic metamorphic episodes in the Appalachians have generally been correlated with the Taconic, Acadian, and Alleghanian orogenies. These events were defined largely on the basis of the stratigraphic record in the adjacent foreland areas. Because of the tectonic resolution now available, however, we believe the simple division of Paleozoic Appalachian orogenesis into the Taconic, Acadian, and Alleghanian is no longer adequate. For example, the Alleghanian orogeny is now known to include a sequence of events that are diachronous and discontinuous along the Appalachian orogenic belt and has been expanded to include a period of >80 Ma (Dallmeyer and others, 1986; Secor and others, 1986; Hatcher and others, 1989). We will therefore not emphasize these divisions and will focus simply on the ages of deformation.

REGIONAL GEOLOGIC SETTING

The Blue Ridge structural province (fig. 1) is bounded on the northwest by the Blue Ridge fault system (Holston-Iron Mountain, Great Smoky, Cartersville) and on the southeast by the Brevard fault zone

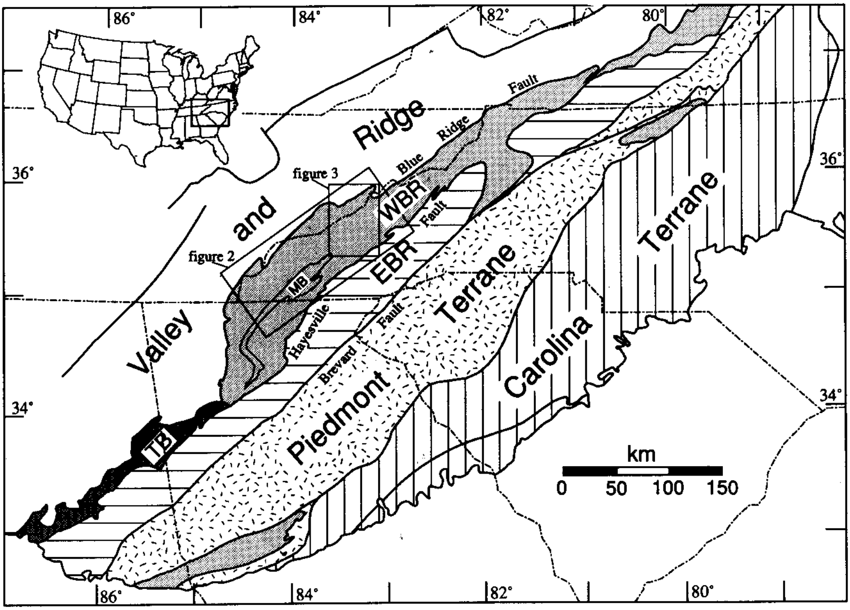


Fig. 1. Regional tectonic map of the southern Appalachian orogen. EBR, eastern Blue Ridge; WBR, western Blue Ridge; MB, Murphy belt; TB, Talladega belt (modified from Hatcher and others, 1990).

(King, 1955; Hatcher, 1972). The Blue Ridge thrust transported imbricated crystalline thrust sheets composed of Precambrian basement, Late Proterozoic-early Paleozoic metasedimentary and metavolcanic rocks, and Paleozoic plutonic units northwestward over Paleozoic sedimentary rocks during Carboniferous-Permian (Alleghanian) orogenesis (Hatcher and others, 1989). Seismic reflection characteristics (Cook and others, 1979; Harris and others, 1981; Çoruh and others, 1987) suggest that most of the Blue Ridge thrust sheet is separated from autochthonous basement by 1 to 5 km of duplicated (duplexed) lower Paleozoic rocks with or without Proterozoic sedimentary rocks or basement. Following emplacement, the Blue Ridge thrust sheet was folded as a result of duplex thrusting within underlying thrust sheets (Boyer and Elliott, 1982; Woodward, 1985; Hatcher, 1991).

The Blue Ridge is divided into two contrasting lithostratigraphic terranes by the Hayesville-Gossan Lead fault (fig. 1), which represents a regional terrane boundary separating rocks initially deposited on North American continental basement from sequences deposited on oceanic or attenuated continental crust of uncertain palinspastic affinities (Hatcher, 1978). This boundary is believed to have been at least locally tectonically active immediately prior to attainment of maximum metamorphic conditions. It was locally reactivated following metamorphism (Hatcher and Goldberg, 1991).

Rocks within the western Blue Ridge comprise a complex sequence of basement gneisses, plutonic units, metasedimentary and metavolcanic rift sequences, together with rifted continental margin and platform successions. Basement rocks consist of polymetamorphic gneisses and associated granitic intrusives that record radiometric ages ranging between 1000 to 1200 Ma (Davis, Tilton, and Wetherill, 1962; Fullagar and Odom, 1973) indicating a Grenville affinity. Nonconformably overlying basement are Late Proterozoic rift-related sequences including metasedimentary and metavolcanic rocks of the Mount Rogers and Grandfather Mountain Formations, and metasedimentary rocks of the Ocoee Supergroup (King and others, 1958; Hadley, 1970). Along the western edge of the Blue Ridge, these Late Proterozoic sequences are overlain both conformably and unconformably by shelf deposits of the Cambrian Chilhowee Group and overlying Shady Dolomite and Rome Formation (Colton, 1970). In eastern segments of the western Blue Ridge, the Ocoee Supergroup is also conformably overlain by the Murphy Group, a diverse sequence of variably metamorphosed clastic and carbonate lithologies (fig. 1).

The structure of the western Blue Ridge is dominated by numerous northwest-vergent thrust faults of contrasting age and character. The earliest thrust faults display ductile fabrics that formed prior to attainment of peak metamorphic conditions, and mylonites developed along these fault zones are typically annealed (Hatcher and Goldberg, 1991). Younger faults, such as those along the frontal Blue Ridge thrust zone, postdate metamorphism and are characterized by predominantly brittle fabrics. Younger faults commonly truncate earlier faults and have been locally reactivated.

The earliest folds recognized in the western Blue Ridge occur in the Great Smoky Mountains area and are east to east-northeast trending and premetamorphic (Hamilton, 1961; Hadley and Goldsmith, 1963). They have been overprinted by one or more generations of more northeasterly-trending ductile folds. The dominant foliation postdates the earliest folding and is commonly crenulated or transposed in more internal portions of the western Blue Ridge.

Rock units exposed in the western Blue Ridge were affected by a progressive Paleozoic Barrovian-type regional metamorphism (Carpenter, 1970). Metamorphic grade (fig. 2) generally increases from the northwest (unmetamorphosed or sub-greenschist facies) to the southeast (at least kyanite grade). The pattern of metamorphism is more complex in the Murphy syncline where grade decreases inward toward the synclinal axis (fig. 2). Retrogressive metamorphic textures have been described throughout the western Blue Ridge. They have generally been attributed to either prolonged cooling following an initial early Paleozoic metamorphism (Dallmeyer, Courtney, and Wooten, 1978) or a distinct later

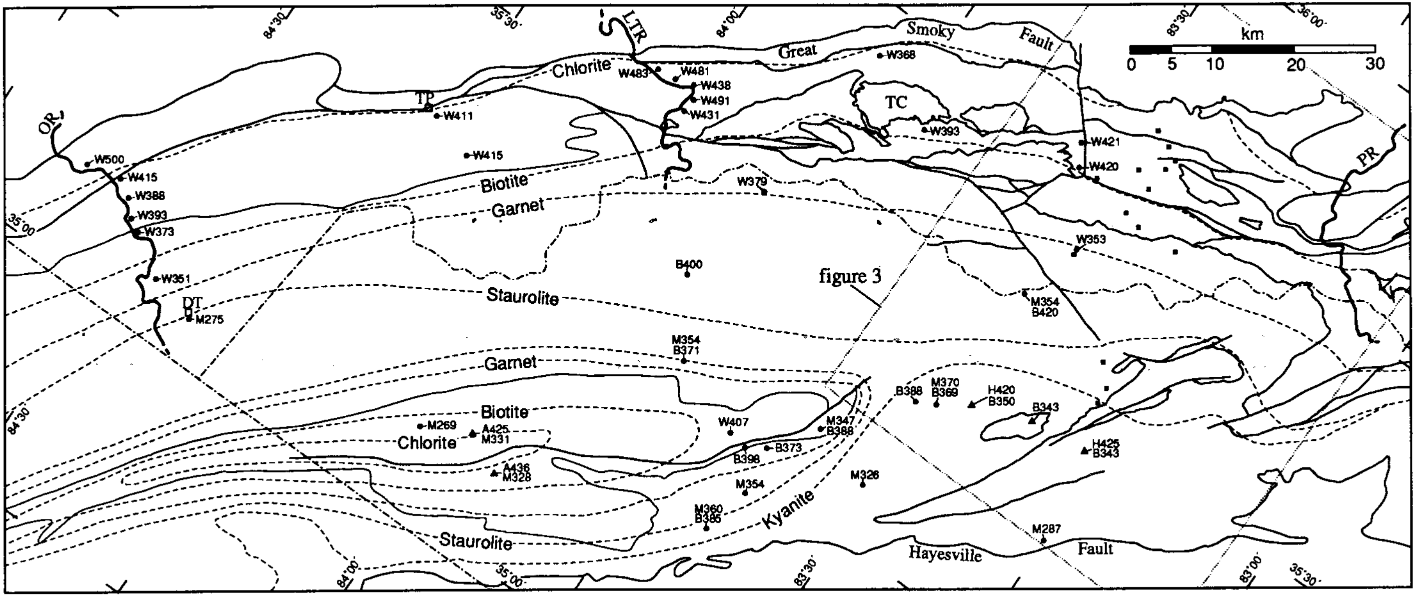


Fig. 2. Generalized geologic map of the southern Appalachian western Blue Ridge in the vicinity of the study area showing the approximate positions of metamorphic isograds and locations and results of previous geochronologic studies (modified from Hardeman, 1966; North Carolina Geological Survey, 1985; Hatcher and Goldberg, 1991). A, amphibole; B, biotite; H, hornblende; M, muscovite; W, whole rock. Circles from K–Ar results compiled in Kish (1991); triangles from $^{40}Ar/^{39}Ar$ results from Dallmeyer (1975, 1988); squares indicate sample locations from the present study. DT, Ducktown; LTR, Little Tennessee River; OR, Ocoee River; PR, Pigeon River; TC, Tuckaleechee Cove; TP, Tellico Plains.

regional metamorphic overprint (Hatcher, 1978; Hatcher and Odom, 1980).

ORDOVICIAN TECTONISM IN THE SOUTHERN APPALACHIANS

The record of Ordovician orogenic activity in eastern North America, now known as the Taconic orogeny, was first described in the Hudson Valley of New York on the basis of an angular unconformity separating folded rocks as young as Middle Ordovician from overlying Early to Late Silurian units. A similar but slightly older orogenic event was also recognized in the southern Appalachians and termed by Kay (1942) and Rodgers (1953) the Blountian phase of the Taconic orogeny. Drake and others (1989) suggested that the Blountian phase was a separate and distinct tectonic event that occurred prior to the Taconic.

Ordovician orogenesis has also been considered important within internal parts of the southern Appalachian orogen (Butler, 1972; Hatcher, 1972, 1978; Dallmeyer, 1975; Butler, 1991). This interpretation was based both on the stratigraphic record in the foreland and geochronologic data reported from internal portions of the orogen. As discussed previously, however, fossil evidence and proposed stratigraphic correlations have questioned the existence of pre-Devonian tectonothermal activity (Tull and Guthrie, 1985; Unrug and Unrug, 1990). Evidence presented both for and against Ordovician orogenic activity in the southern Appalachian western Blue Ridge is discussed below.

Evidence for Ordovician Tectonic Activity in the southern Appalachians

The foreland stratigraphic record.—Kay (1942) first recognized the significance of extensive early Middle Ordovician clastic rocks in the western Valley and Ridge province (Blountian clastic wedge) and suggested that these sediments were derived from an eastern highland that had developed during the Blountian disturbance. This clastic wedge was deposited in the Sevier basin of eastern Tennessee and overlies a major unconformity developed at the top of the Lower Ordovician-Upper Cambrian Knox Group. The clastic sequence includes shallow shelf, debris flows, pelagic, turbidite, and shallow water-subaerial deposits (Shanmugan, 1980).

Coarse polymictic conglomerates occur within turbiditic sequences of the Blountian wedge that contain fragments derived from most of the Lower Paleozoic units stratigraphically down to and including the Chilhowee Group (Kellberg and Grant, 1956; Lowry, 1972; Mack, 1985). This has been interpreted to suggest substantial structural relief adjacent to the basin had developed prior to deposition. An easterly source is indicated for the conglomerates and associated turbidite deposits (Shanmugan, 1980). Orogenic activity prior to conglomerate deposition is also indicated by the internal deformation and metamorphism displayed within some of the rock fragments within the conglomerates. Kellberg and Grant (1956) recognized that fragments of Lower Cambrian clastics were "altered from sandstone to vitreous quartzite." Mack (1985) described metapelitic slate and phyllite fragments within sandstones of the Blountian clastic wedge indicating a low-grade metamorphic source.

Geochronologic evidence.—Available radiometric ages for the western Blue Ridge in the vicinity of the study area are summarized in figure 2. Summaries of geochronology available in this area may be found in Glover and others (1983), Drake and others (1989), Osberg and others (1989), Butler (1991), and Kish (1991). Initial geochronologic studies in the Blue Ridge recognized a limited area of early Paleozoic (450 Ma) metamorphism within lower grade rocks exposed within the western Blue Ridge, and a more extensive area of middle Paleozoic (350 Ma) metamorphism was documented within higher grade parts of the southern Appalachians (Long, Kulp, and Eckelman, 1959; Kulp and Eckelman, 1961). These authors concluded that two distinct early Paleozoic regional metamorphic events were recorded. The younger was believed to have culminated at 350 Ma and appeared to have variably rejuvenated micas that had initially crystallized at 460 Ma. More recent interpretations proposed that the younger ages may date post-metamorphic uplift and cooling through appropriate closure temperatures; they therefore represent only minimum ages for metamorphism (Hadley, 1964; Armstrong, 1966; Butler, 1972). The 350 Ma ages have, therefore, commonly been interpreted to record regionally diachronous cooling following a single phase of tectonothermal activity at 450 to 480 Ma.

Rb-Sr whole-rock results were reported by Fullagar and Bottino (1970) from Ducktown, Tennessee, and Ore Knob, North Carolina. These results indicated that Paleozoic metamorphism in the Blue Ridge occurred prior to 475 Ma. Kish, Fullagar, and Dabbagh (1976) reported 440 ± 13 Ma Rb-Sr whole-rock ages for unmetamorphosed pegmatites exposed near Bryson City, North Carolina, and ages of approx 400 Ma were listed for pegmatites from the Spruce Pine district of North Carolina. These dates were interpreted to reflect minimum dates for regional metamorphism in the western Blue Ridge.

Kish and Harper (1973) reported conventional K-Ar whole-rock ages for slate and phyllite from the western Blue Ridge (northwest of the biotite isograd). These ranged from 420 to 400 Ma and were interpreted to indicate that regional metamorphism occurred at $\geq$400 Ma ago. A wide range of K-Ar whole-rock ages for lower greenschist facies slates from the western Great Smoky Mountains were grouped at approx 480 Ma and at 430 to 370 Ma (Kish, 1982). The relatively wide range in the younger group of K-Ar ages was interpreted to reflect either slow cooling and/or partial rejuvanation by a 390 Ma recrystallization event.

Dallmeyer (1975) reported $^{40}Ar/^{39}Ar$ ages for hornblende and biotite from recrystallized basement rocks from the western Blue Ridge in North Carolina. Hornblende concentrates displayed internally concordant spectra which defined plateau ages of 420 to 425 Ma (recalculated from the initially published results using the decay constants proposed by Steiger and Jäger, 1977). Biotite concentrates yielded plateau ages of 350 to 360 Ma. The spectra provided no indication of extraneous argon contamination and were interpreted by Dallmeyer (1975) to date times of post-metamorphic cooling through the contrasting temperatures required for intracrystalline retention of argon. The hornblende dates were considered to represent a minimum age for the high-grade Paleozoic metamorphism recorded in the area and were used to develop a post-metamorphic thermal model that suggested attainment of peak metamorphic conditions at 480 Ma.

Dallmeyer (1988) reported $^{40}Ar/^{39}Ar$ data on amphibole and muscovite concentrates from rocks within the Murphy belt. Amphibole results suggested post-metamorphic cooling through argon closure temperatures at 425 to 440 Ma. Discordant age spectra were interpreted to result from a distinct later thermal overprint that locally effected partial argon loss from amphibole at 325 to 350 Ma. The overprint was of sufficient magnitude to rejuvenate completely muscovite within interlayered pelitic schist which records plateau ages of approx 330 Ma. These results were interpreted by Dallmeyer (1988) to indicate a polymetamorphic evolution that involved both Ordovician and Carboniferous thermal events.

Evidence precluding Ordovician Orogenic activity in the southern Appalachians

Fossil evidence.—Unrug and Unrug (1990) reported a fossil assemblage within regionally metamorphosed rocks of the Walden Creek Group (Ocoee Supergroup) exposed along westernmost sections of the Blue Ridge in Tennessee. This assemblage is reported as trilobite, ostracode, bryozoan, and microcrinoid fragments, and agglutinated foraminifera. The foraminiferal assemblage was interpreted to be Silurian or younger. More recently, Unrug, Unrug, and Palmes (1991) reported additional fossil assemblages from the Walden Creek Group that include calcispheres and calcareous foraminifera. These fossils were interpreted to reflect a Late Devonian (Frasnian) to earliest Mississippian age, suggesting that the previously assumed Late Proterozoic depositional age for the Walden Creek Group, and possibly the entire Ocoee, be abandoned (Unrug, Unrug, and Palmes, 1992). In addition, the new paleontologic age assignment questions the geologic significance of most previously reported geochronology from the western Blue Ridge (Dallmeyer, 1975; Kish, Fullagar, and Dabbagh, 1976; Dallmeyer, 1988). A Late Devonian to earliest Mississippian age for the Walden Creek Group requires that recorded metamorphism must be Carboniferous or younger and leaves only a remote possibility that some of the earliest (premetamorphic) deformation occurred during Devonian orogenesis.

The validity and interpretation of the fossils described by Unrug and Unrug (1990) and Unrug, Unrug, and Palmes (1991) have been questioned because of: (1) inconsistencies with previously documented stratigraphic relationships (King and others, 1958; Keller, ms); (2) other fossil evidence (Knoll and Keller, 1979); and (3) lack of confirmation by other workers (Broadhead and Hatcher, 1992). A conformable contact between the Sandsuck Formation (uppermost unit of the Walden Creek Group) and overlying Chilhowee Group argues against a middle Paleozoic age for the Walden Creek Group, because an Early Cambrian age for the Chilhowee Group has been clearly documented (Walcott, 1890; Laurence and Palmer, 1963; Simpson and Sundberg, 1987; Walker and Driese, 1991). Although stratigraphic relations between the Chilhowee Group and upper portions of the Walden Creek Group are equivocal in the Great Smoky Mountains (King and others, 1958; Hamilton, 1961), stratigraphic relations immediately east of the Great Smoky Mountains are certain. Detailed mapping in this area (Ferguson and Jewell, 1951; Keller, ms) has documented that the Walden Creek Group (including the Sandsuck Formation) stratigraphically underlies the Chilhowee Group (fig. 3). A continuous, possibly unconformable, succession from the Sandsuck Formation to the Chilhowee Group is also exposed northeast and within and adjacent to the Hot Springs window (Oriel, 1951; Bearce, 1969; Walker and Simpson, 1991). Stratigraphic evidence, therefore, appears to preclude a post-Early Cambrian age for the Walden Creek Group.

The detailed toxonomic identification of fossils presented by Unrug and Unrug (1990) and Unrug, Unrug, and Palmes (1991) has been questioned (Rodgers, 1991; Broadhead, Hatcher and Costello, 1991). Because of their small size and poor preservation, it has been suggested that these fossils cannot be confirmed as different from Ordovician and Cambrian fossils. For example, similar foraminifera have been reported from Lower Cambrian rocks of west Africa (Culver and others, 1990; Culver, 1991). Rodgers (1991) suggested that the fossils described by Unrug and Unrug (1990) may belong to the Tommotian stage, the lowest known fossiliferous stage of the Cambrian. In addition, nonfossil allochems similar to the calcareous foraminifera and calcispheres reported by Unrug, Unrug, and Palmes (1991) have been described by Broadhead and Hatcher (1992) from a carbonate horizon within the Sandsuck Formation that clearly underlies Early Cambrian rocks of the Chilhowee Group along the Parksville reservoir. These authors suggested that the structures described by Unrug, Unrug, and Palmes (1991) within other units of the Walden Creek Group may be peripherally micritized and diagenetically recrystallized inorganic grains. Broadhead, Hatcher and Costello (1991), however, suggested that the bryozoan and crinoid fragments pictured in Unrug and Unrug (1990) and Unrug, Unrug, and Palmes (1991) are not known before the Middle Ordovician.

Other problems arising from a middle to late Paleozoic depositional age suggested for the Walden Creek Group include: (1) lack of megafossils within carbonate units interpreted to represent platform margin deposits (Rodgers, 1991; Broadhead, Hatcher, and Costello, 1991); and (2) absence of conodonts that are common microfossils in Silurian and Devonian carbonate rocks formed in a wide variety of facies (Broadhead, Hatcher, and Costello, 1991). A middle to late Paleozoic age assignment also conflicts with other fossil evidence from the Walden Creek Group. Knoll and Keller (1979) suggested a late Precambrian age on the basis of

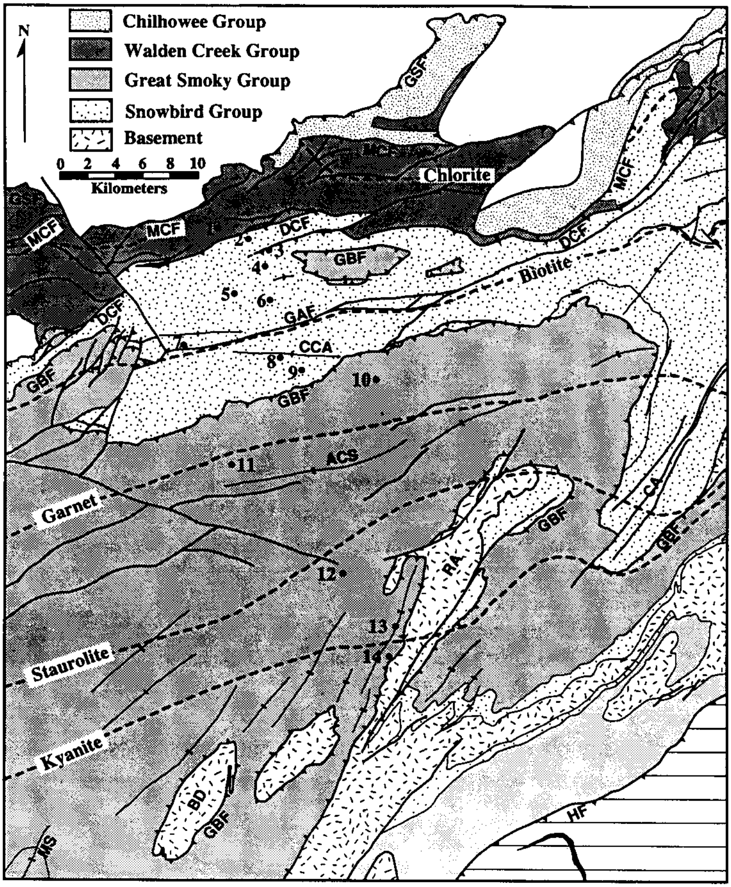


Fig. 3. Geologic map of the eastern Great Smoky Mountains area showing sample localities and metamorphic isograds. ACS, Alum Cave syncline; BD, Bryson dome; CA, Cataloochee anticlinorium; CCA, Copeland Creek anticline; DCF, Dunn Creek fault; GAF, Gatlinburg fault; GBF, Greenbrier fault; GSF, Great Smoky fault; HF, Hayesville fault; MCF, Miller Cove fault; MS, Murphy syncline; RA, Ravensford anticline (modified from King, Neumann, and Hadley, 1968).

acritarchs, particularly *Bavlinella faveolata* (Shepleva); a species now known to range into the Cambrian (Knoll and Swett, 1985). Broadhead, Hatcher and Costello (1991) recently relocated and described soft-bodied metazoan macrofossils originally discovered by Phillips (ms) in the Sandsuck Formation. This confirms that the Sandsuck Formation can be no younger than Early Cambrian. Thus, as indicated by Broadhead, Hatcher, and Costello (1991), evaluating the true geologic significance of the reports of Unrug and Unrug (1990) and Unrug, Unrug, and Palmes (1991) awaits both independent confirmation and careful taxonomic investigation of the fossils in order to determine the biostratigraphic significance of these fossils.

Stratigraphic evidence.—Several workers have suggested a correlation between the Murphy belt (western Blue Ridge) and nearby Talladega belt (Crickmay, 1936; Stose and Stose, 1944; Hadley, 1980; Tull and Guthrie, 1985). The Talladega belt, situated in the southwest part of the western Blue Ridge, occupies a structurally equivalent position between the foreland thrust belt and the eastern Blue Ridge (fig. 1). Fossiliferous Lower Devonian strata within the Talladega belt have been regionally metamorphosed to lower greenschist facies. Whole-rock conventional K-Ar dates for slate from the Talladega belt yielded a mean age of 399 ± 17 Ma (Kish, 1990). These are compatible with paleontologic controls that indicate Early to Middle Devonian metamorphism in the Talledega belt.

Tull and Guthrie (1985) and Tull and Groszos (1988) have argued that uppermost stratigraphic levels of the Murphy Group (Mineral Bluff Formation) are correlative with lithologically similar clastic sequences exposed in the Talladega belt (Talladega Group). Upper portions of the Talladega Group (upper Lay Dam Formation and Jemison Chert) are paleontologically restricted to the Silurian-Early Devonian (Tull and others, 1988). In addition, these authors proposed that a major unconformity separates the uppermost Mineral Bluff Formation from underlying formations of the Murphy Group. This unconformity has been correlated with a similar post-Early Ordovician unconformity in the Talladega belt. This unconformity was suggested to have developed prior to metamorphism and significant deformation because of the low angle and continuity of structural fabrics below the unconformity into overlying higher-grade units of the Murphy belt rocks and Great Smoky Group (Dallmeyer, Courtney, and Wooten, 1978; Tull and Groszos, 1988). If these stratigraphic correlations are correct, regional metamorphism in the western Blue Ridge must have been post-Silurian.

Despite gross lithologic similarities between rocks of the Talladega Group and Mineral Bluff Formation (Tull and Groszos, 1988), other correlations are possible. An alternative possibility suggested by Tull and Groszos (1988) is that deposition of the Mineral Bluff clastic sequence pre-dated the Talladega Group. This would require that the Talladega Group be missing within the Murphy belt. Based on analogies with known deep-water stratigraphic successions exposed in other areas, Thomas and Hatcher (1988) suggested that the Murphy succession may represent deep-water proximal deposits of the Middle Ordovician Blountian clastic wedge. Similarly, Hatcher, Costello, and Broadhead (1992) suggested that the Murphy belt sequence could represent a more distal facies of the platform sequence that was deposited later (time-transgressively) eastward. In this interpretation, the clastic sequence above the proposed unconformity in the Murphy belt is part of the Blountian clastic wedge. A post-Ordovician age for rocks of the Murphy belt has therefore not been stratigraphically demonstrated.

GEOLOGY OF THE STUDY AREA

Stratigraphy

The present study area is located in the eastern Great Smoky Mountains of the western Blue Ridge (figs. 1, 3). The dominant stratigraphic unit exposed in this area is the Ocoee Supergroup, which consists of a 12 to 15 km thick sequence of predominantly clastic metasedimentary rocks (King and others, 1958; Hadley, 1970; Rast and Kohles, 1986). Numerous thrust faults are present in the study area affecting both basement rocks and units within the overlying Ocoee Supergroup. Although faulting locally complicates regional stratigraphic relations, the Ocoee Supergroup has been divided into three major and contrasting lithologic sequences. In ascending stratigraphic order, these include the Snowbird Group, the Great Smoky Group, and the Walden Creek Group (King and others, 1958). The Snowbird Group is generally considered the oldest because locally it nonconformably overlies basement and is itself conformably overlain by the Great Smoky Group. The Great Smoky Group, however, also locally nonconformably overlies basement (Hadley and Goldsmith, 1963). The Walden Creek Group conformably succeeds both the Snowbird Group (Ferguson and Jewell, 1951; Keller, ms) (fig. 3) and the Great Smoky Group (Hurst and Schlee, 1962; Hernon, 1964; Costello and Hatcher, 1986, 1991) in different areas.

Fault Systems

The Great Smoky Mountains area records the effects of at least two major deformational events that have produced five distinct fault systems (Woodward and others, 1991). These include the Great Smoky, Gatlinburg, Miller Cove, Dunn Creek, and Greenbrier fault systems (fig. 3). The Great Smoky and Gatlinburg systems are regarded as late Paleozoic in age. The Dunn Creek and Greenbrier fault systems are premetamorphic and are believed to have been active in the early Paleozoic. An additional early Paleozoic thrust sheet, now floored by the Miller Cove fault, was emplaced during regional cleavage development (Connelly and Woodward, 1992). Early Paleozoic structures may be distinguished from late Paleozoic structures on the basis of their relations to regional cleavage and metamorphic isograds and on the ductile nature of the early fault fabrics. Predominantly brittle fabrics are characteristic of late Paleozoic faults in this area.

The Great Smoky thrust fault forms the northwestern boundary of the western Blue Ridge province in the vicinity of the study area and is part of the Blue Ridge thrust system. The Great Smoky fault is dominantly brittle, and internal deformation of late Paleozoic age within the Great Smoky thrust sheet in this area is minor.

The younger Gatlinburg fault system is a primarily east-northeast trending system of brittle, high-angle structures (fig. 3). Both dip-slip and strike-slip motion are recognized; however, dip slip predominates in the study area. Maximum displacement is approximately 2000 m (King, 1964). Although King (1964) described these faults as thrusts, Woodward and others (1991) noted that normal separations also occur. Woodward and others (1991) interpreted these faults as "late" structures related to folding of the Great Smoky thrust sheet because they parallel the post-emplacement folding.

The predominantly brittle Miller Cove fault branches from the Great Smoky fault (fig. 3) and separates cleaved and metamorphosed rocks of the Ocoee Supergroup from generally uncleaved and unmetamorphosed rocks of the upper Ocoee Supergroup, Chilhowee Group, and younger strata (Costello, ms; Hatcher and others, 1989). The Miller Cove fault locally represents the frontal Blue Ridge fault where the branch line between the Great Smoky and Miller Cove faults has been eroded. Pre-late Paleozoic cleavage, folds, and ductile thrust faults within the Miller Cove thrust sheet, however, suggest that this thrust sheet was at least locally active prior to the late Paleozoic (Connelly and Woodward, 1992).

The Greenbrier fault is a folded, low-angle thrust separating rocks of the Great Smoky Group from the underlying Snowbird Group (fig. 3). A horizontal displacement of at least 24 km has been estimated based on stratigraphic criteria (Hadley and Goldsmith, 1963) and structural reconstructions (Connelly and Woodward, 1922). Because the Greenbrier fault does not offset metamorphic isograds or affect regional cleavage, it has been considered a premetamorphic structure (Hadley and Goldsmith, 1963; King, 1964; Milton, 1983).

The Dunn Creek thrust fault separates Walden Creek Group footwall rocks from Snowbird Group hanging wall rocks throughout most of the foothills area (fig. 3). A premetamorphic age for the Dunn Creek fault is indicated by truncation of this fault by later synmetamorphic thrust faults (Connelly and Woodward, 1992) and lack of offset of the chlorite isograd northeast of the study area (Keller, ms).

Folds

The earliest post-Grenville folds recognized in the Great Smoky Mountains include the east-trending F_1 Cartertown-Copeland Creek anticline (Dunn Creek thrust sheet) and the Alum Cave syncline (Greenbrier thrust sheet) (fig. 3). They are transected by and predate a regional S_1 foliation (Hamilton, 1961; Connelly and Woodward, 1992). These folds have been described as truncated by the Greenbrier fault and thus were interpreted to predate emplacement. Connelly and Woodward (1992), however, reinterpreted them as rootless, ramp-related folds that formed during emplacement of the Greenbrier and Dunn Creek thrust sheets. They suggested that a foreland-style thrust belt existed prior to overprinting by regional cleavage, metamorphism, and ductile folding.

Early folds and faults were affected by a second generation of folds (F_2) which exhibit axial-planar S_1 cleavage and include most of the mesoscopic and some map-scale folds in the Miller Cove and Dunn Creek thrust sheets. Some folds are cored by ductile thrust faults that locally truncate premetamorphic faults (Connelly and Woodward, 1992). Within the Miller Cove thrust sheet, cleavage is axial planar to folds (Witherspoon, ms; Sack, ms) and is commonly parallel to the ductile thrust faults. Within the Dunn Creek thrust sheet, however, cleavage transects most east-trending folds and is axial planar to northeast-trending second-generation folds. Interference between these two fold generations occurs throughout the Dunn Creek thrust sheet and results in steeply plunging second-generation folds with axial-planar cleavage. Connelly and Woodward (1992) suggested that these F_2 folds and the S_1 regional cleavage formed during pre-late Paleozoic movement of the present Miller Cove thrust sheet.

Third-generation (F_3) folds in this area trend north-northeast and include the Ravensford anticline, Cataloochee anticlinorium, and many smaller associated folds which occur primarily within the Greenbrier thrust sheet (fig. 3). These are the second-generation folds described by Hadley and Goldsmith (1963). Larger F_3 anticlines locally expose basement rocks. These folds are typically open but may be isoclinal, and most are asymmetric with axial planes dipping southeastward (Hadley and Goldsmith, 1963). Folding of regional S_1 foliation by F_3 folds indicates that they postdated regional metamorphism (Hadley and Goldsmith, 1963; Witherspoon, ms). S_2 crenulation cleavage is parallel to axial planes of F_3 folds (Hadley and Goldsmith, 1963; Mohr, 1973). Nonpenetrative S_2 cleavage is locally present north of the Greenbrier fault and becomes progressively more penetrative south of the garnet isograd. At higher metamorphic grades, S_2 cleavage becomes the dominant foliation (Hadley and Goldsmith, 1963).

Metamorphism

Isograds in the eastern Great Smoky Mountains suggest a progressive Barrovian-type metamorphism that generally increases from subchlorite grade (within frontal units) to at least kyanite grade (fig. 3). In detail, however, the metamorphic history of this area is complex. Hadley and Goldsmith (1963) recognized that the peak of thermal metamorphism was both preceded and followed by deformation and a lower-grade metamorphism. Petrographic evidence suggests that S_1 regional cleavage and schistosity formed during an early kinematic phase of metamorphism. This was followed by a higher-grade, static recrystallization that involved formation of porphyroblasts of chloritoid, biotite, garnet, staurolite, and/or kyanite. Porphyroblasts are randomly oriented and overgrow all preexisting foliations (Hadley and Goldsmith, 1963; King, 1964). Following porphyroblast growth, an S_2 crenulation cleavage locally formed and fractured, offset, and rotated porphyroblasts. Minor quartz and chlorite formed in pressure shadows adjacent to some porphyroblasts (Hadley and Goldsmith, 1963).

Mohr (1973) described similar textural characteristics within garnet and lower metamorphic grades west of Bryson dome (fig. 3). Above garnet grade, however, snowball textures suggest that porphyroblasts grew during development of S_2. One or more post-S_2 minor retrograde events are indicated by the local presence of chlorite and white mica selvages over kyanite crystals and local pseudomorphic replacement of staurolite and kyanite by sericite and/or chlorite (Mohr, 1973). Textures indicative of polymetamorphism were also described by Power and Forrest (1971) within the Murphy belt. Early foliation was overgrown by garnet. Subsequent folding was followed by growth of sillimanite, staurolite, and biotite. Garnets are commonly included within staurolite.

ANALYTICAL METHODS

Muscovite concentrates were prepared from three samples of argillaceous Thunderhead metasandstone collected within high-grade portions of the western Blue Ridge in Tennessee and North Carolina (staurolite and kyanite zones). Eleven slate/phyllite samples were collected for whole-rock analysis from lower grade areas along an approximately north-south (across strike) transect. Lithologies included the Walden Creek, Snowbird, and Great Smoky Groups. Sample localities are indicated in figure 3. Location coordinates of the dated samples are provided in app. 1.

Illite Crystallinity

The seven slate/phyllite samples from the chlorite zone were crushed and sieved following wire-brush removal of weathered surfaces and thorough washing. They were prepared for determination of illite crystallinity by desegregation in a shatter box for 20 sec. Bulk <2 μm size-fractions were isolated by differential settling in Atterberg cylinders and centrifugation following techniques listed in Reuter (1985). Illite crystallinity of the <2 μm size fractions was determined from oriented sedimentation slides by comparison of the (001) and (100) quartz (internal standard) reflections following the methods of Weber (1972). Cross-calibration of 28 samples (correlation coefficient = 0.97) from Reuter (1985, 1987) suggests that the following boundary values are appropriate for the equipment setting employed at the University of Georgia (compared with the calibrations of Teichmüller, Teichmüller, and Weber, 1979): greenschist/anchizone = 115; anchizone/diagenesis = 350. In the present study, boundaries between the upper anchizone/middle anchizone and middle anchizone/lower anchizone are defined at crystallinity values of 190 and 270 respectively (fig. 4). According to Kubler (1967) and Dunoyer de Segonzac (1969, 1970), minimum illite crystallinity values are reached within the epizone whereas Teichmüller, Teichmüller, and Weber (1979) define the greenschist/anchizone boundary by the first appearance of minimum crystallinity values. As a result, rocks suggested to reflect epizonal metamorphism according to Kubler (1967) and/or Dunoyer de Segonzac (1969, 1970) are classified as upper anchizone by Teichmüller, Teichmüller, and Weber (1979; fig. 4).

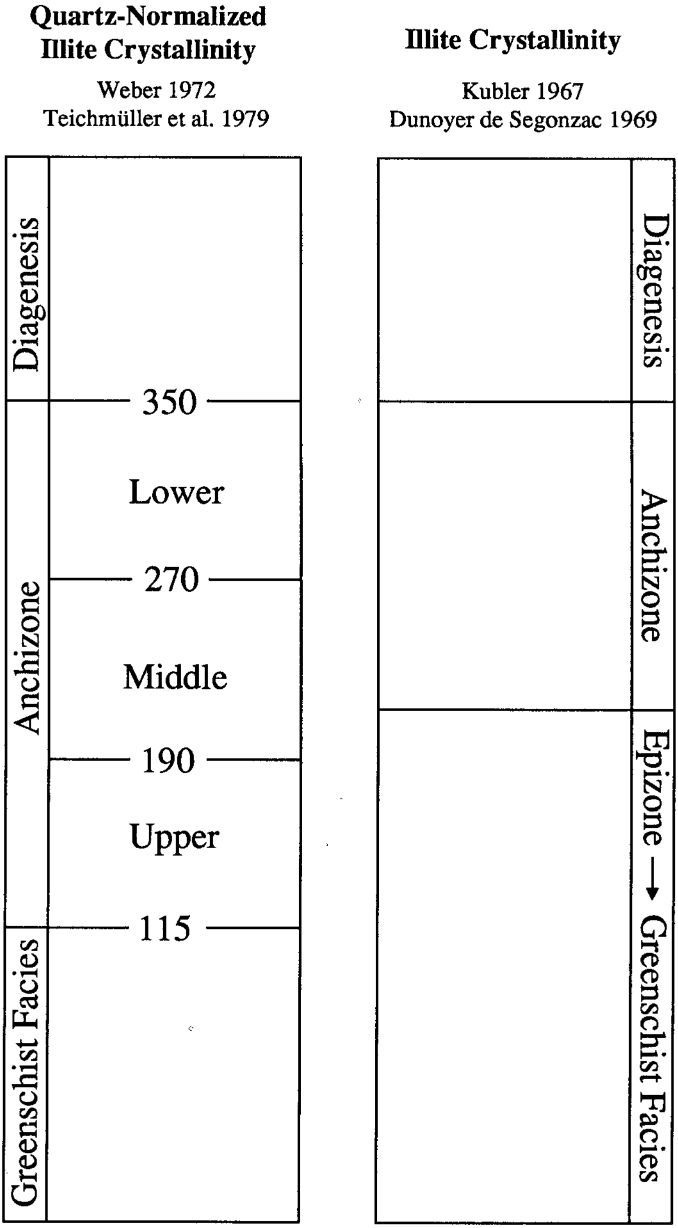


Fig. 4. Comparison of illite crystallinity based on the contrasting methods of Kubler (1967) and Teichmüller, Teichmüller, and Weber (1979).

$^{40}Ar/^{39}Ar$ Analyses

Techinques used during $^{40}Ar/^{39}Ar$ analyses generally followed those described in detail by Dallmeyer and Gil-Ibarguchi (1990). The whole-rock slate/phyllite samples were prepared by crushing and sieving to 80/100 mesh followed by thorough washings. Muscovite concentrates (>99 percent) were prepared from crushed and sized rock powders using flotation and magnetic separation. Mineral concentrates and whole-rock powder were wrapped in aluminum-foil packets, encapsulated in sealed quartz vials, and irradiated for 40 hrs at the TRIGA reactor at the U.S. Geological Survey in Denver, Colorado. Variations in the flux of neutrons along the length of the irradiation assembly were monitored with several mineral standards, including MMhb-1 (Samson and Alexander, 1987). The samples were incrementally heated until fusion in a double-vacuum, resistance-heated furnace. Measured isotopic ratios were corrected for total system blanks and the effects of mass discrimination. Interfering isotopes produced during irradiation were corrected using factors reported by Dalrymple and others (1981) for the TRIGA reactor. Apparent $^{40}Ar/^{39}Ar$ ages were calculated from corrected isotopic ratios using the decay constants and isotopic abundance ratios listed by Steiger and Jäger (1977) following the methods described in Dallmeyer and Keppie (1987).

Intralaboratory uncertainties have been calculated by statistical propagation of uncertainties associated with measurement of each isotopic ratio (at two standard deviations of the mean) through the age equation. Interlaboratory uncertainties are ±1.25 to 1.5 percent of the quoted age. Total-gas ages have been computed for each sample by appropriate weighting of the age and percentage ^{39}Ar released within each temperature increment. A "plateau" is considered to be defined if the ages recorded by two or more contiguous gas fractions (with similar apparent K/Ca ratios) each representing >4 percent of the total ^{39}Ar evolved (and together constituting >50 percent of the total quantity of ^{39}Ar evolved) are mutually similar within a ±1 percent intralaboratory uncertainty. Analysis of the MMhb-1 monitor indicates that apparent K/Ca ratios may be calculated through the relationship 0.518 (±0.005) × ($^{39}Ar/^{37}Ar$) corrected.

RESULTS

Illite Crystallinity

The quartz-normalized illite crystallinity indices for the bulk <2 μm size fractions isolated from the slate/phyllite samples collected within the chlorite zone range between 116 and 436 (table 1). The results correlate with metamorphic conditions ranging from lower greenschist facies to the diagenesis zone (following the classification of Teichmüller, Teichmüller, and Weber, 1979; fig. 4).

TABLE 1

Quartz-normalized illite crystallinity determined on bulk <2 μm size fractions at the locations sampled for $^{40}Ar/^{39}Ar$ dating in the chlorite zone of the Eastern Great Smoky Mountains, Western Blue Ridge

Sample	Crystallinity
1	436
2	195
3	187
4	147
5	138
6	116
7	144

Note. Comparison of (001) reflection in illite and (100) reflection in internal quartz standard (after Weber, 1972).

$^{40}Ar/^{39}Ar$

Whole-rock slate/phyllite.—Eleven whole-rock slate/phyllite samples from the western Blue Ridge were analyzed with ^{40}Ar/^{39}Ar incremental-release techniques. The analytical data are listed in table 2 and portrayed as age spectra in figures 5 and 6.

Seven whole-rock samples (1–7) have been analyzed from the chlorite zone (diagenesis to uppermost anchizone/lowermost greenschist). These display variably discordant ^{40}Ar/^{39}Ar spectra (fig. 5) that define total-gas ages ranging between 499 and 389 Ma. In general, only minor intrasample variation in the apparent ^{40}Ar/^{39}Ar ages is recorded by most intermediate-temperature gas fractions. Considerable intersample variation, however, exists in the intermediate-temperature ages. For the lowest grade sample (1: diagenesis) these are approx 500 Ma. For samples from the middle-upper achizone (2–5) they are approx 450 Ma (plateaux are defined for samples 3–5). For samples from the upper anchizone/lowermost greenschist facies (6–7), intermediate temperature ages are 380 to 420 Ma. The intermediate temperature fractions are all characterized by similar intrasample apparent K/Ca ratios (fig. 5) indicating that experimental evolution of gas occurred from compositionally uniform populations of intracrystalline "sites." These are interpreted to correspond with constituent, very fine-grained white mica. Low-temperature gas increments are characterized by systematically increasing apparent ages and variable apparent K/Ca ratios. Petrographic characteristics suggest that these increments correspond to experimental gas evolution from variably rejuvenated detrital potassium feldspar and chlorite. Systematically decreasing apparent K/Ca ratios are displayed by most high-temperature increments and are interpreted to reflect experimental evolution of gas from detrital plagioclase feldspar. In the lower grade samples (1–3: diagenesis and middle/upper anchizone), increasingly

TABLE 2

$^{40}Ar/^{39}Ar$ analytical data for incremental heating experiments on whole-rock slate/phyllite samples from the western Blue Ridge, Tennessee-North Carolina

Release Temp. (°C)	$(^{40}Ar/^{39}Ar)^*$	$(^{36}Ar/^{39}Ar)^*$	$(^{37}Ar/^{39}Ar)^C$	^{39}Ar % of total	^{40}Ar non-atm.$^+$	$^{36}Ar_{Ca}$ %	Apparent Age (Ma)**
Diagenesis-Lower Anchizone							
Sample 1.	J = 0.010522						
425	18.04	0.00563	0.013	6.59	90.76	0.06	286.8 ± 2.2
450	21.34	0.00126	0.018	6.88	98.24	0.39	359.5 ± 2.7
475	27.37	0.00099	0.028	5.71	98.92	0.78	452.2 ± 2.0
505	29.92	0.00059	0.024	14.03	99.40	1.09	491.1 ± 2.0
535	30.79	0.00074	0.026	13.33	99.28	0.97	503.0 ± 1.3
565	30.75	0.00079	0.023	13.59	99.22	0.79	502.2 ± 1.2
590	30.89	0.00084	0.025	13.18	99.18	0.81	503.9 ± 1.4
615	31.85	0.00067	0.024	8.05	99.36	0.97	518.4 ± 1.2
635	33.71	0.00033	0.026	5.23	99.70	2.11	546.2 ± 1.8
650	36.31	0.00051	0.032	4.10	99.58	1.69	581.5 ± 2.7
675	39.75	0.00070	0.036	3.06	99.47	1.41	627.4 ± 3.9
705	43.75	0.00099	0.035	2.40	99.32	0.95	679.2 ± 4.5
735	49.27	0.00167	0.036	2.12	98.99	0.59	747.2 ± 4.3
Fusion	53.60	0.00306	0.044	1.73	98.31	0.39	795.6 ± 4.3
Total	30.83	0.00114	0.025	100.00	98.65	0.91	499.3 ± 1.5
Middle-Upper Anchizone							
Sample 2.	J = 0.010302						
425	10.93	0.00594	0.017	3.58	83.90	0.08	162.8 ± 3.2
450	11.81	0.00176	0.022	4.89	95.56	0.34	198.3 ± 2.6
475	18.19	0.00139	0.020	4.01	97.72	0.39	303.4 ± 1.8
505	27.32	0.00052	0.028	7.95	99.43	1.48	445.0 ± 1.9
535	28.35	0.00063	0.033	7.78	99.34	1.42	459.5 ± 2.2
565	27.99	0.00061	0.027	11.07	99.34	1.19	454.4 ± 1.6
590	27.24	0.00052	0.026	9.80	99.42	1.37	443.9 ± 1.9
615	26.65	0.00020	0.027	11.25	99.76	3.57	436.7 ± 2.1
635	26.34	0.00042	0.029	11.29	99.51	1.84	431.1 ± 1.9
655	26.38	0.00017	0.029	6.45	99.79	4.47	432.8 ± 2.0
675	26.63	0.00046	0.029	6.26	99.47	1.68	435.2 ± 1.3
705	27.30	0.00011	0.032	4.59	99.87	7.69	446.5 ± 2.1
735	28.57	0.00089	0.036	3.49	99.07	1.10	461.5 ± 3.8
770	30.24	0.00111	0.038	2.28	98.91	0.94	484.5 ± 3.7
815	31.41	0.00038	0.059	3.14	99.64	4.24	504.1 ± 2.9
Fusion	33.51	0.00066	0.148	2.15	99.44	6.10	532.3 ± 3.3
Total	25.84	0.00076	0.031	100.00	98.67	2.21	420.6 ± 1.8
Sample 3.	J = 0.009965						
425	11.60	0.00595	0.022	3.07	84.81	0.10	168.7 ± 1.9
450	12.24	0.00123	0.013	3.54	97.00	0.28	201.8 ± 1.8
475	18.80	0.00039	0.015	3.09	99.37	1.08	308.0 ± 2.1
505	25.11	0.00060	0.041	4.95	99.28	1.86	400.3 ± 1.6
535	27.18	0.00036	0.042	8.18	99.60	3.19	430.9 ± 0.8
56µ	28.45	0.00048	0.045	9.16	99.49	2.50	448.2 ± 2.2
590	28.65	0.00066	0.045	7.03	99.32	1.87	450.4 ± 1.8
615	28.42	0.00045	0.042	11.88	99.52	2.55	447.9 ± 1.6
635	28.39	0.00043	0.041	10.59	99.55	2.62	447.7 ± 2.1
655	28.36	0.00029	0.039	10.46	99.69	3.72	447.8 ± 0.9
675	28.02	0.00029	0.037	9.65	99.69	3.53	443.0 ± 2.0
700	28.12	0.00037	0.038	7.15	99.60	2.82	444.1 ± 2.1
730	29.89	0.00023	0.033	4.41	99.77	4.01	469.4 ± 1.5
765	32.45	0.00021	0.049	3.39	99.80	6.41	504.5 ± 2.0
800	35.82	0.00087	0.094	2.07	99.29	2.92	547.3 ± 2.2
Fusion	34.66	0.00069	0.233	1.38	99.45	9.20	532.7 ± 2.3
Total	27.16	0.00061	0.042	100.00	99.01	2.85	428.0 ± 1.2
Total without 425-535 °C, 730 °C-fusion				65.92			447.0 ± 1.1

Upper Anchizone

Sample 4.	J = 0.009662						
375	8.58	0.00857	0.043	0.23	70.44	0.14	102.3 ± 6.3
400	9.48	0.00309	0.002	1.67	90.32	0.01	143.4 ± 4.0
425	12.55	0.00137	0.003	2.46	96.74	0.05	200.1 ± 2.7
450	13.23	0.00064	0.040	2.89	98.55	1.70	214.1 ± 1.6
475	18.57	0.00084	0.045	3.08	98.65	1.47	293.9 ± 1.8
500	22.40	0.00081	0.046	4.91	98.92	1.54	349.9 ± 1.3
525	28.39	0.00027	0.082	4.54	99.72	8.29	436.2 ± 2.0
545	30.37	0.00043	0.070	5.93	99.58	4.48	462.4 ± 1.9
565	30.49	0.00039	0.071	6.55	99.62	4.91	464.1 ± 2.2
585	30.21	0.00043	0.065	8.50	99.58	4.09	460.2 ± 1.9
605	30.01	0.00047	0.074	8.09	99.54	4.30	457.4 ± 2.1
625	29.97	0.00051	0.062	8.93	99.49	3.30	456.6 ± 2.2
645	29.82	0.00033	0.056	9.54	99.67	4.68	455.4 ± 1.8
670	29.50	0.00021	0.055	9.21	99.78	7.03	451.4 ± 2.1
695	29.75	0.00042	0.053	6.97	99.58	3.46	454.0 ± 1.6
720	29.88	0.00041	0.052	5.20	99.59	3.41	455.8 ± 1.2
745	30.50	0.00066	0.057	3.26	99.36	2.33	463.3 ± 2.2
775	31.36	0.00009	0.050	2.61	99.91	15.54	477.0 ± 2.6
810	31.76	0.00057	0.065	3.48	99.47	3.13	480.5 ± 1.9
Fusion	31.27	0.00055	0.087	1.95	99.49	4.29	474.1 ± 2.6
Total	28.01	0.00053	0.059	100.00	99.22	4.31	428.3 ± 1.5
Total without 375-500 °C, 745 °C-fusion				68.91			457.2 ± 1.3

Sample 5.	J = 0.009953						
500	16.00	0.00180	0.043	5.56	96.66	0.65	258.3 ± 1.2
540	26.88	0.00046	0.059	6.44	99.49	3.50	425.7 ± 1.5
570	27.38	0.00022	0.062	7.73	99.76	7.79	433.8 ± 1.6
590	27.48	0.00022	0.057	9.69	99.76	7.08	435.1 ± 1.3
610	27.50	0.00005	0.063	6.19	99.95	37.23	436.1 ± 1.6
630	27.57	0.00015	0.055	11.12	99.84	10.13	436.7 ± 0.9
660	27.66	0.00033	0.059	13.44	99.64	4.85	437.3 ± 1.1
690	27.87	0.00013	0.055	10.73	99.86	11.54	441.0 ± 1.1
720	28.05	0.00013	0.056	9.53	99.86	11.55	443.5 ± 1.3
750	28.47	0.00009	0.057	5.51	99.90	16.85	449.6 ± 1.0
780	28.79	0.00043	0.067	2.90	99.56	4.27	452.7 ± 2.6
830	29.08	0.00004	0.082	4.51	99.96	55.78	458.3 ± 1.8
900	29.09	0.00109	0.097	5.01	98.90	2.42	454.2 ± 2.1
Fusion	28.98	0.00027	0.373	1.63	99.81	37.31	456.4 ± 3.1
Total	27.20	0.00034	0.065	100.00	99.56	12.27	430.2 ± 1.1
Total without 500-540 °C				68.44			437.8 ± 0.9

Upper Anchizone-Lower Greenschist

Sample 6.	J = 0.010095						
500	20.49	0.00186	0.034	5.64	97.30	0.50	330.9 ± 1.7
540	25.22	0.00062	0.048	12.78	99.27	2.11	406.4 ± 1.3
570	25.68	0.00029	0.049	8.28	99.65	4.52	414.6 ± 1.5
600	25.75	0.00020	0.048	17.56	99.77	6.59	416.0 ± 2.2
630	26.04	0.00028	0.046	11.07	99.68	4.53	419.9 ± 2.5
660	26.35	0.00023	0.048	9.93	99.73	5.58	424.5 ± 1.5
690	26.69	0.00038	0.041	8.56	99.57	2.98	428.8 ± 2.4
720	26.71	0.00040	0.042	6.06	99.55	2.83	429.0 ± 2.0
760	27.15	0.00065	0.042	5.27	99.28	1.74	434.1 ± 3.0
800	27.38	0.00071	0.050	3.78	99.23	1.92	437.3 ± 3.6
860	27.33	0.00009	0.049	5.00	99.90	15.49	439.1 ± 2.5
920	28.04	0.00041	0.064	5.21	99.56	4.25	448.0 ± 3.2
Fusion	29.44	0.00057	0.213	0.87	99.46	10.16	467.2 ± 4.2
Total	25.98	0.00045	0.048	100.00	99.46	4.56	418.0 ± 2.1

TABLE 2
(continued)

Sample 7.	J = 0.010111						
450	17.21	0.00240	0.044	4.34	95.87	0.50	278.2 ± 2.0
500	23.63	0.00027	0.054	4.74	99.66	5.47	385.3 ± 1.8
540	23.38	0.00026	0.047	16.00	99.66	4.87	381.6 ± 1.7
570	23.38	0.00009	0.047	12.00	99.88	14.40	382.4 ± 2.1
600	23.71	0.00031	0.044	7.22	99.60	3.80	386.3 ± 2.2
630	23.68	0.00023	0.043	8.32	99.70	5.00	386.1 ± 2.1
660	23.93	0.00032	0.047	8.74	99.60	4.07	389.5 ± 2.0
690	24.04	0.00028	0.043	8.19	99.64	4.13	391.2 ± 2.3
720	24.22	0.00022	0.050	6.84	99.73	6.32	394.2 ± 2.6
750	24.40	0.00036	0.049	7.72	99.56	3.75	396.3 ± 2.0
780	24.92	0.00039	0.063	5.10	99.53	4.39	403.6 ± 2.1
840	25.64	0.00087	0.069	4.81	98.99	2.16	412.1 ± 2.9
900	26.81	0.00033	0.147	4.23	99.66	12.07	431.4 ± 2.5
Fusion	32.60	0.00138	1.416	1.74	99.08	28.00	510.3 ± 2.3
Total	23.90	0.00040	0.077	100.00	99.46	6.22	388.5 ± 1.7

Biotite Zone

Sample 8.	J = 0.010470						
450	49.25	0.14428	0.136	1.80	13.44	0.03	120.8 ± 7.2
500	18.51	0.00674	0.027	6.60	89.22	0.11	287.8 ± 1.6
540	21.23	0.00212	0.035	8.55	97.03	0.45	352.3 ± 1.2
565	21.13	0.00105	0.034	8.43	98.52	0.87	355.7 ± 0.8
590	20.76	0.00094	0.034	6.86	98.65	0.99	350.5 ± 1.3
615	20.53	0.00088	0.032	6.79	98.72	0.99	347.1 ± 2.6
645	20.41	0.00080	0.032	9.37	98.83	1.08	345.7 ± 0.9
675	20.52	0.00072	0.036	6.41	98.95	1.35	347.7 ± 1.0
710	20.67	0.00021	0.044	9.71	99.69	5.65	352.4 ± 1.2
745	20.95	0.00056	0.089	7.23	99.21	4.32	355.2 ± 1.2
780	21.23	0.00049	0.129	5.87	99.33	7.12	359.9 ± 1.1
815	21.51	0.00045	0.094	7.13	99.38	5.65	364.4 ± 1.0
860	22.31	0.00065	0.115	10.10	99.15	4.79	375.8 ± 1.0
Fusion	23.61	0.00097	0.207	5.13	98.83	5.78	394.4 ± 1.2
Total	21.52	0.00380	0.067	100.00	96.68	2.91	348.8 ± 1.3
Sample 9.	J = 0.010291						
450	14.02	0.00734	0.019	1.77	84.50	0.07	207.5 ± 4.8
500	17.56	0.00666	0.022	1.72	88.76	0.09	268.3 ± 7.4
550	23.52	0.00223	0.012	3.37	97.17	0.14	381.0 ± 4.5
580	23.10	0.00092	0.003	2.65	98.79	0.10	380.5 ± 3.3
620	22.93	0.00066	0.006	3.88	99.12	0.25	379.2 ± 2.6
650	22.41	0.00034	0.006	5.71	99.53	0.50	372.8 ± 2.0
680	22.10	0.00073	0.007	5.29	99.00	0.26	366.3 ± 1.1
710	21.82	0.00024	0.005	9.40	99.65	0.58	364.3 ± 0.9
740	21.68	0.00011	0.006	8.54	99.82	1.54	362.7 ± 1.9
770	21.83	0.00074	0.005	9.11	98.97	0.19	362.1 ± 1.3
800	21.99	0.00033	0.006	7.30	99.53	0.51	366.3 ± 1.3
830	22.29	0.00073	0.014	4.44	99.01	0.54	369.2 ± 1.3
860	22.35	0.00007	0.018	5.39	99.88	6.67	373.0 ± 1.7
890	22.61	0.00013	0.012	8.08	99.80	2.34	376.7 ± 1.4
920	23.13	0.00005	0.011	7.95	99.92	6.65	384.8 ± 1.4
980	24.04	0.00036	0.012	10.14	99.53	0.88	397.1 ± 1.6
Fusion	29.12	0.00057	0.105	5.26	99.43	5.02	470.5 ± 1.8
Total	22.62	0.00067	0.014	100.00	98.96	1.76	374.1 ± 1.8

Sample 10.	J = 0.010162						
450	17.00	0.00230	0.018	1.97	95.97	0.21	276.7 ± 3.6
500	21.17	0.00057	0.020	0.93	99.18	0.93	348.9 ± 3.8
540	21.29	0.00128	0.020	0.79	98.20	0.42	347.5 ± 7.0
565	21.32	0.00151	0.015	0.94	97.89	0.27	346.9 ± 2.0
590	21.26	0.00082	0.017	1.57	98.83	0.55	349.0 ± 2.6
615	21.11	0.00040	0.015	1.95	99.42	1.03	348.7 ± 2.8
645	20.95	0.00045	0.015	3.45	99.64	0.89	346.0 ± 1.8
675	20.78	0.00026	0.015	4.32	99.60	1.56	344.2 ± 1.2
710	20.57	0.00019	0.015	8.21	99.70	2.10	341.4 ± 0.7
740	20.38	0.00010	0.015	9.81	99.83	3.92	339.0 ± 0.7
780	20.33	0.00013	0.014	10.85	99.79	2.98	338.1 ± 0.7
815	20.46	0.00010	0.015	12.32	99.83	3.84	340.1 ± 0.6
850	20.76	0.00024	0.014	10.72	99.64	1.65	344.2 ± 0.9
920	21.22	0.00011	0.019	20.84	99.83	4.78	351.7 ± 0.4
Fusion	22.31	0.00009	0.038	11.33	99.87	11.82	368.1 ± 0.8
Total	20.86	0.00023	0.018	100.00	99.64	4.01	345.6 ± 0.9
			Garnet Zone				
Sample 11.	J = 0.010162						
450	38.02	0.04739	0.023	1.05	63.16	0.01	379.6 ± 2.8
500	23.20	0.00738	0.028	0.88	90.58	0.10	336.3 ± 4.1
550	20.37	0.00201	0.028	1.73	97.06	0.38	318.0 ± 1.8
600	19.96	0.00047	0.025	5.58	99.29	1.48	318.7 ± 1.4
640	20.43	0.00018	0.027	8.06	99.72	4.03	326.9 ± 1.5
670	21.71	0.00043	0.025	7.99	99.40	1.58	344.5 ± 1.4
700	21.93	0.00127	0.019	13.38	98.26	0.40	344.1 ± 1.1
730	21.53	0.00083	0.015	8.62	98.84	0.51	340.1 ± 1.6
760	21.29	0.00014	0.013	9.71	99.78	2.41	339.6 ± 0.9
790	21.26	0.00060	0.011	8.27	99.14	0.48	337.2 ± 0.8
820	21.46	0.00007	0.012	6.74	99.88	4.61	342.4 ± 1.0
850	21.50	0.00007	0.011	5.53	99.88	4.24	342.9 ± 1.2
885	21.46	0.00056	0.013	7.93	99.21	0.61	340.3 ± 0.6
920	21.41	0.00021	0.015	3.26	99.68	1.87	341.0 ± 1.5
980	21.57	0.00036	0.014	8.42	99.49	1.05	342.8 ± 1.1
Fusion	23.12	0.00050	0.065	2.87	99.35	3.50	364.6 ± 1.2
Total	21.58	0.00108	0.018	100.00	98.80	1.76	339.7 ± 1.2
Total without 450-600 °C and fusion				79.84			341.6 ± 1.0

* measured.
c corrected for post-irradiation decay of ^{37}Ar (35.1 day 1/2-life).
+ [^{40}Ar tot. − (^{36}Ar atm.) (295.5)]/$^{40}Ar_{tot.}$
**calculated using correction factors of Dalrymple and others (1981); two sigma, intralaboratory errors.

older apparent ages are recorded in the high-temperature increments, which likely relate to source ages. In the higher-grade samples (4–7: upper anchizone/lowermost greenschist facies), the high-temperature increase in apparent ages is much less marked, likely reflecting extensive Paleozoic metamorphic rejuvenation of the detrital plagioclase grains.

Four whole-rock phyllite samples (8–11) have been analyzed from the biotite and garnet zones. These display apparent K/Ca spectra with characteristics generally similar to those described for the lower grade slate/phyllite whole-rock samples (fig. 6); however, there is generally

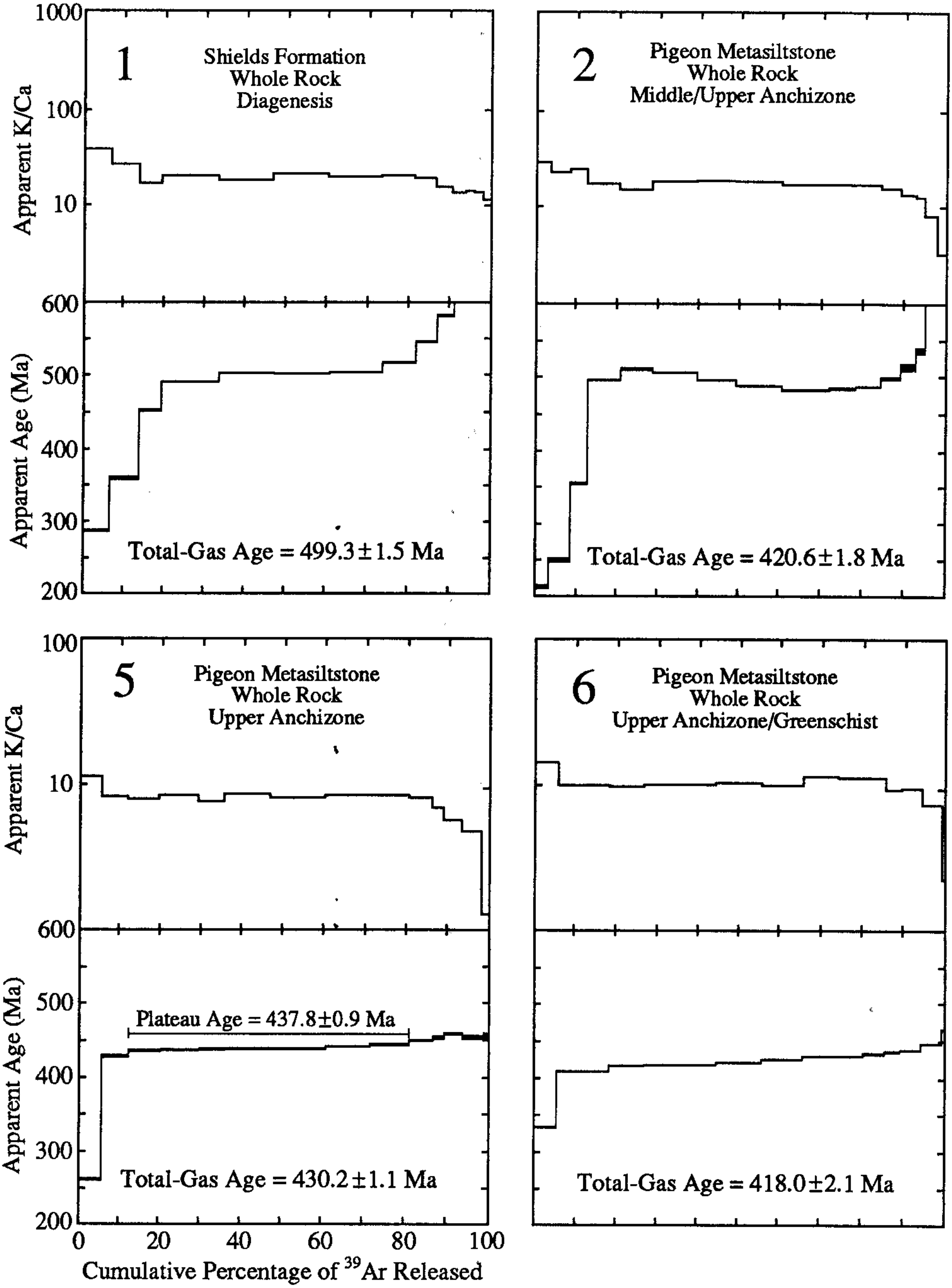
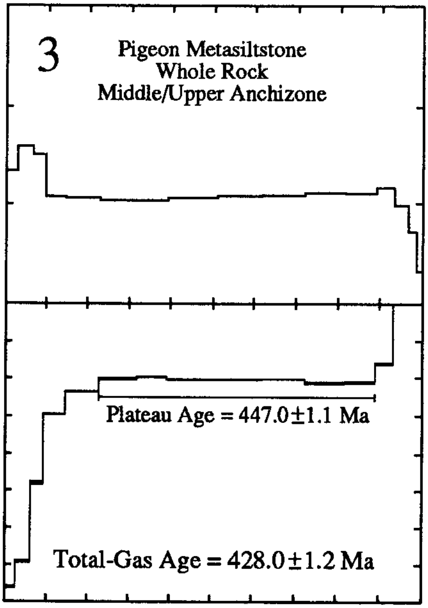
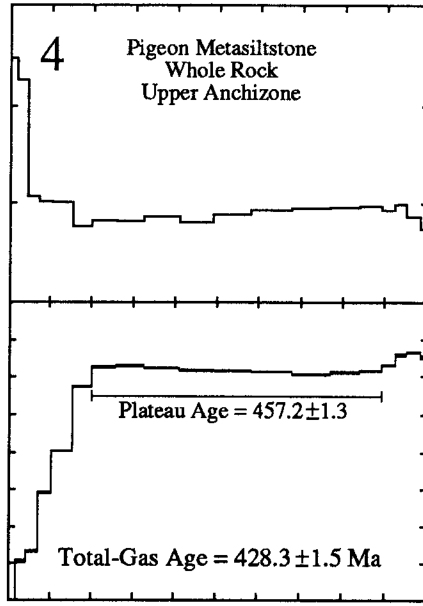

1
Shields Formation
Whole Rock
Diagenesis
Apparent K/Ca
Apparent Age (Ma)
Total-Gas Age = 499.3±1.5 Ma
2
Pigeon Metasiltstone
Whole Rock
Middle/Upper Anchizone
Total-Gas Age = 420.6±1.8 Ma
5
Pigeon Metasiltstone
Whole Rock
Upper Anchizone
Plateau Age = 437.8±0.9 Ma
Total-Gas Age = 430.2±1.1 Ma
6
Pigeon Metasiltstone
Whole Rock
Upper Anchizone/Greenschist
Total-Gas Age = 418.0±2.1 Ma
Cumulative Percentage of 39Ar Released

Figure 5





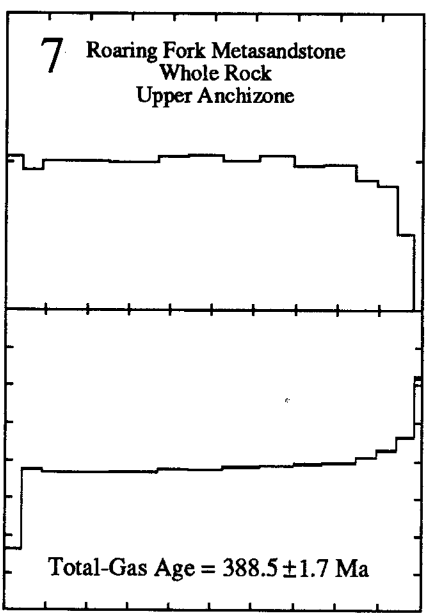


Fig. 5. $^{40}Ar/^{39}Ar$ apparent age and apparent K/Ca spectra of whole-rock analyses of slate/phyllite samples from the chlorite zone, Miller Cove and Dunn Creek thrust sheets. Sample locations shown in figure 3. Analytical uncertainties (two sigma intralaboratory) are represented by vertical width of bars. Experimental temperatures increase from left to right.

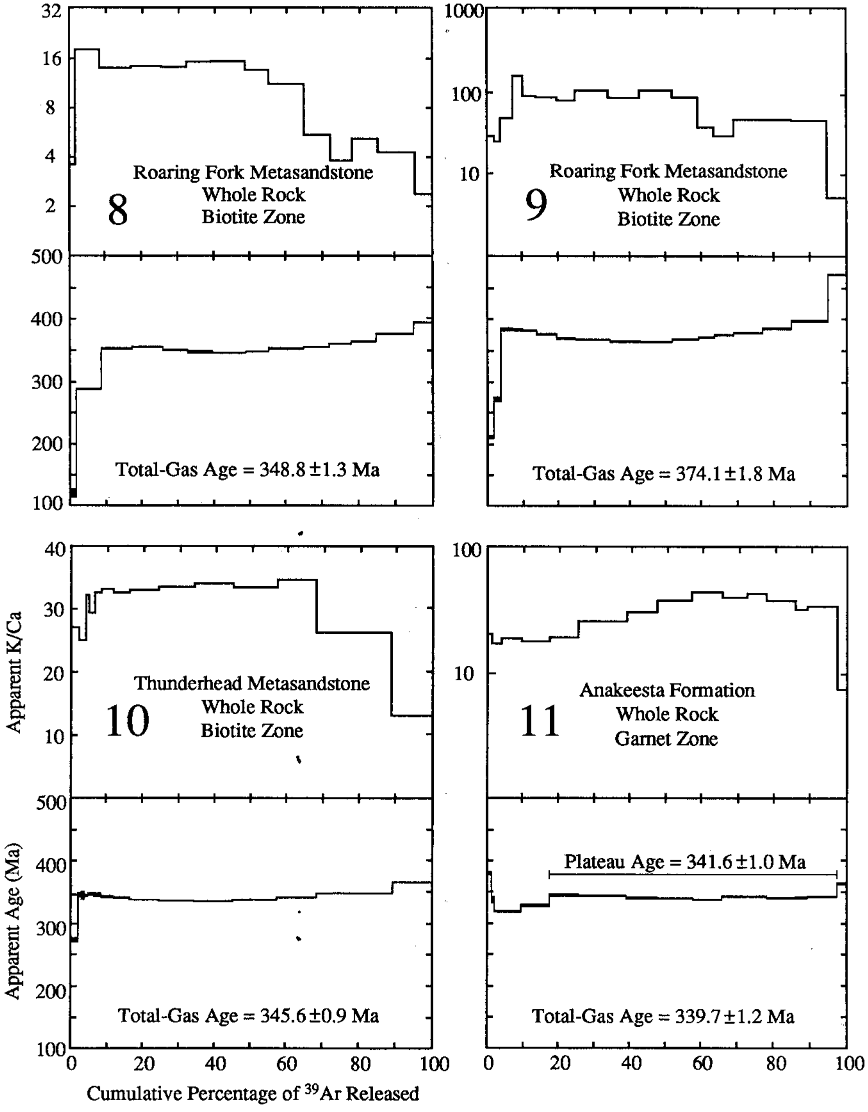


Fig. 6. $^{40}Ar/^{39}Ar$ apparent age and apparent K/Ca spectra of whole-rock analyses of slate/phyllite samples from the biotite and garnet zones, Dunn Creek and Greenbrier thrust sheets. Sample locations shown in figure 3. Data plotted as in figure 5.

much less variation in the apparent ages recorded by intermediate-temperature gas fractions. Most range between 340 and 350 Ma.

Muscovite.—Three muscovite concentrates were analyzed with $^{40}Ar/^{39}Ar$ incremental release techniques. The analytical data are listed in table 3 and are presented as age spectra in figure 7. These samples display only slightly discordant $^{40}Ar/^{39}Ar$ spectra (fig. 7) that define plateau ages of 362 Ma (12), 372 Ma (13), and 377 Ma (14). Apparent K/Ca ratios are very large and display no significant or systematic intrasample variations. Therefore, they are not presented with the age spectra. The plateau ages are interpreted to date the last cooling through temperatures required for intracrystalline retention of argon. Although not fully calibrated experimentally, using the preliminary data of Robbins (ms) in the diffusion equations of Dodson (1973) suggests that temperatures of 350 ± 25°C are required for argon retention in muscovite.

COMPARISON WITH PREVIOUS GEOCHRONOLOGY

The $^{40}Ar/^{39}Ar$ results from the present study area are similar to and compatible with results from previous geochronological studies in adjacent areas of the western Blue Ridge. Whole-rock conventional K-Ar ages determined for Ocoee Supergroup slates and phyllites from the chlorite, biotite, and garnet metamorphic zones (Kish and Harper, 1973; Kish, 1982, 1991) yielded age patterns that are similar to those of the present study. Samples from a transect along the Little Tennessee River (chlorite and biotite zones) yielded ages that range from 483 to 379 Ma and show a gradual younging to the southeast (fig. 2). Samples from a transect along the Ocoee River (chlorite and biotite zones) yield somewhat younger ages between 415 and 351 Ma (fig. 2). Two samples collected near Tellico Plains (chlorite zone) yield ages of 411 and 415 Ma, whereas samples from north of Tuckaleechee cove (chlorite zone) show slightly younger ages of 393 and 368 Ma (fig. 2). Two samples from several kilometers west of the present transect (chlorite zone) yield ages of 420 and 421 Ma (fig. 2). A sample from Mt. LeConte (garnet zone) yielded an age of 353 Ma. This result is comparable to the plateau age of 342 Ma determined for sample 11 also collected from Mt. LeConte (fig. 2).

Muscovite ages from the present study are similar to most previously determined muscovite ages from the western Blue Ridge which range from 350 to 370 Ma (fig. 2). The 377 Ma age determined for sample 14, however, is the oldest muscovite age in this part of the western Blue Ridge and is believed to date closely peak metamorphic conditions in this area. $^{40}Ar/^{39}Ar$ muscovite ages from the Murphy belt area are approx 330 Ma (Dallmeyer, 1988; fig. 2) and somewhat younger than other areas of the western Blue Ridge. $^{40}Ar/^{39}Ar$ biotite ages from several kilometers south of the muscovite samples of the present study (fig. 2) yield $^{40}Ar/^{39}Ar$ ages of 350 to 360 Ma (Dallmeyer, 1975; recalculated from the initially published results using the decay constants proposed by Steiger and Jäger, 1977). These slightly younger ages are likely the result of the lower temperature required for Ar retention in biotite. Most other biotite ages from the Blue Ridge are anomalously old compared to muscovite ages from the same sample suggesting the presence of excess ^{40}Ar in biotite in this area (Kish, 1989).

TABLE 3

$^{40}Ar/^{39}Ar$ analytical data for incremental heating experiments on muscovite concentrates from the western Blue Ridge, Tennessee-North Carolina

Release temp. (°C)	$(^{40}Ar/^{39}Ar)$*	$(^{36}Ar/^{39}Ar)$*	$(^{37}Ar/^{39}Ar)^c$	^{39}Ar % of total	^{40}Ar non-atm.+	$^{36}Ar_{Ca}$ %	Apparent Age (Ma)**
			Staurolite Zone				
Sample 12.	J = 0.009750						
550	19.63	0.01226	0.026	1.10	81.53	0.06	261.6 ± 7.9
600	22.29	0.00248	0.007	3.31	96.69	0.008	344.1 ± 2.7
630	22.18	0.00116	0.004	2.87	98.43	0.09	348.2 ± 2.0
660	22.02	0.00009	0.003	2.58	99.85	0.97	350.3 ± 1.6
690	21.99	0.00119	0.006	3.66	98.38	0.13	345.2 ± 2.5
720	22.14	0.00059	0.005	3.70	99.19	0.22	350.0 ± 2.4
750	22.48	0.00124	0.006	5.10	98.34	0.13	352.1 ± 1.6
780	22.91	0.00128	0.007	7.53	98.33	0.15	358.2 ± 1.1
810	23.13	0.00091	0.003	12.00	98.82	0.10	363.0 ± 0.6
840	22.65	0.00083	0.004	9.89	98.89	0.12	356.3 ± 0.9
870	22.64	0.00096	0.005	8.56	98.72	0.14	355.6 ± 0.9
900	22.89	0.00062	0.005	9.43	99.17	0.22	360.7 ± 1.0
950	23.31	0.00067	0.004	17.94	99.12	0.17	366.5 ± 0.6
Fusion	23.35	0.00034	0.006	12.33	99.55	0.46	368.6 ± 0.5
Total	22.83	0.00097	0.005	100.00	98.69	0.21	358.3 ± 1.1
Total without 550-750 °C				77.69			362.3 ± 0.7
Sample 13.	J = 0.009751						
550	22.31	0.00931	0.013	1.38	87.65	0.04	314.7 ± 5.9
600	24.91	0.00174	0.005	1.69	97.91	0.07	384.9 ± 5.5
630	24.22	0.00130	0.009	2.62	98.39	0.18	376.8 ± 5.2
660	23.34	0.00083	0.005	3.89	98.92	0.17	366.3 ± 1.7
690	22.97	0.00086	0.008	4.99	98.87	0.26	360.8 ± 1.7
720	23.65	0.00110	0.007	6.63	98.60	0.16	369.5 ± 1.4
750	24.08	0.00099	0.006	12.52	98.76	0.16	376.1 ± 0.6
780	23.90	0.00079	0.006	11.96	99.00	0.19	374.4 ± 0.7
810	23.63	0.00109	0.008	5.85	98.61	0.19	369.4 ± 1.9
840	23.65	0.00096	0.005	6.76	98.78	0.13	370.2 ± 1.4
870	23.64	0.00077	0.007	5.97	99.02	0.24	370.9 ± 1.2
900	23.40	0.00082	0.005	6.58	98.95	0.16	367.2 ± 1.4
940	23.46	0.00088	0.003	8.20	98.87	0.09	367.8 ± 1.4
980	23.68	0.00075	0.005	11.50	99.05	0.17	371.5 ± 0.8
Fusion	23.73	0.00047	0.005	9.47	99.39	0.28	373.4 ± 0.6
Total	23.68	0.00099	0.006	100.00	98.73	0.18	370.5 ± 1.3
Total without 550-690 °C				85.44			371.7 ± 0.9
			Kyanite Zone				
Sample 14.	J = 0.009281						
550	25.98	0.04631	0.051	0.79	47.31	0.03	194.9 ± 10.2
600	25.62	0.00683	0.014	2.39	92.10	0.06	357.2 ± 6.8
630	26.78	0.00238	0.017	2.60	97.35	0.20	390.8 ± 3.1
660	25.93	0.00218	0.003	3.06	97.49	0.04	380.2 ± 2.1
690	25.20	0.00170	0.009	4.32	97.99	0.15	372.2 ± 2.6
720	25.16	0.00197	0.010	4.85	97.66	0.13	370.6 ± 1.3
750	25.77	0.00168	0.005	9.91	98.05	0.09	380.0 ± 2.0
780	25.81	0.00181	0.005	7.39	97.91	0.07	380.1 ± 1.4
810	25.71	0.00112	0.004	11.82	98.69	0.11	381.5 ± 0.6
840	25.51	0.00148	0.005	7.45	98.26	0.08	377.2 ± 1.7
870	25.33	0.00159	0.005	7.61	98.12	0.08	374.4 ± 1.9
900	25.21	0.00138	0.005	7.89	98.36	0.09	373.5 ± 2.0
950	25.22	0.00142	0.003	9.05	98.31	0.06	373.6 ± 1.9
1000	25.25	0.00110	0.003	7.19	98.69	0.08	375.3 ± 1.9
1050	25.52	0.00085	0.004	8.31	99.00	0.14	380.0 ± 1.7
Fusion	25.63	0.00052	0.005	5.39	99.68	0.25	382.8 ± 1.8
Total	25.53	0.00191	0.006	100.00	97.78	0.10	375.8 ± 0.8
Total without 550-600 °C				91.16			377.2 ± 0.7

* measured.

^{c} corrected for post-irradiation decay of ^{37}Ar (35.1 day 1/2-life).

\+ $[^{40}Ar\ tot. - (^{36}Ar\ atm.)\ (295.5)]/^{40}\ tot.$

** calculated using correction factors of Dalrymple and others (1981); two sigma, intralaboratory errors.

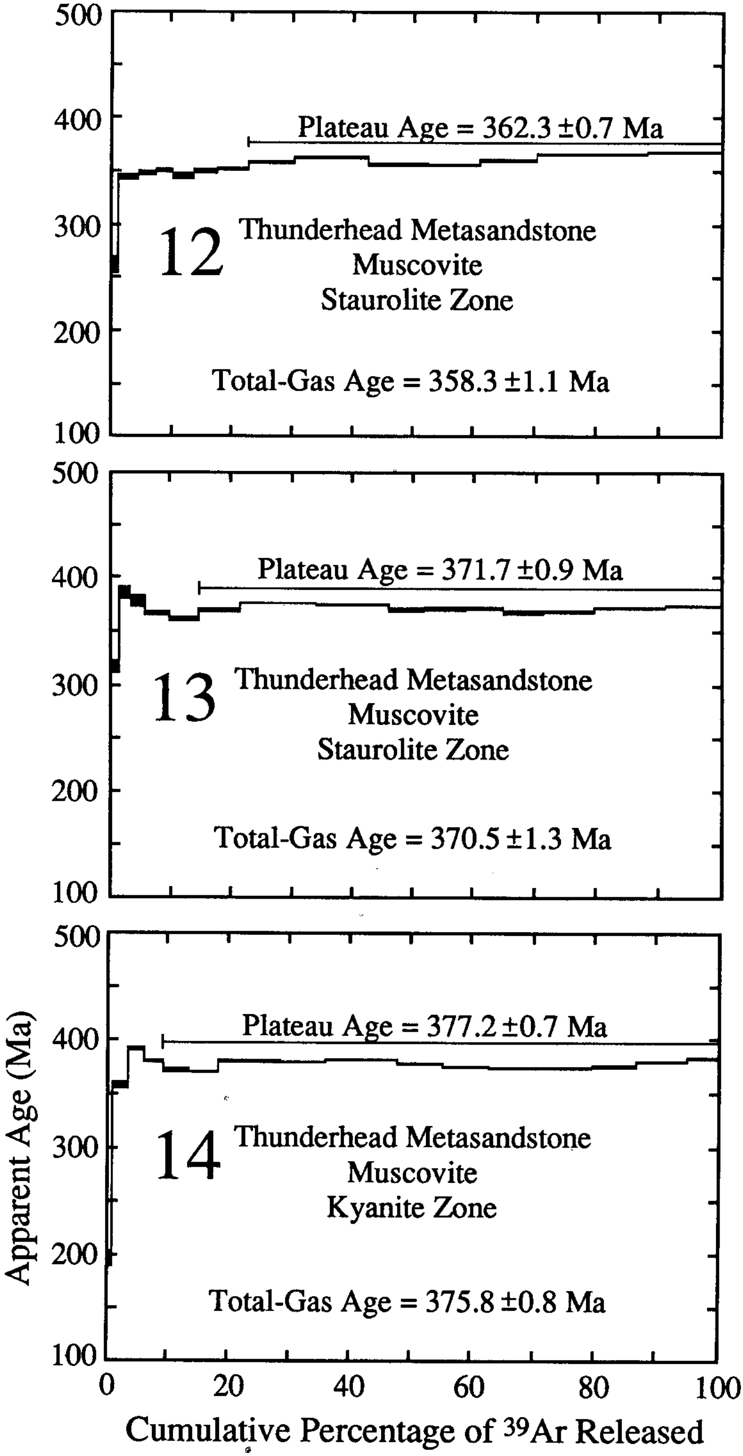


Fig. 7. $^{40}Ar/^{39}Ar$ apparent age spectra of muscovite concentrates from the staurolite and kyanite zones, Greenbrier thrust sheet. Sample locations shown in figure 3. Data plotted as in figure 5.

INTERPRETATION

Whole-rock slate/phyllite samples from the middle-upper anchizone (2–5) yield $^{40}Ar/^{39}Ar$ plateau and intermediate temperature ages of 450 Ma. These dates contrast with apparent ages of 350 Ma defined by intermediate-temperature increments of samples from the biotite and garnet zones (8–11). It is unlikely that samples collected only several kilometers apart could have experienced cooling histories after a single thermal event that would result in an 100 Ma difference in apparent ages. Results from the present study are therefore not consistent with previously reported models for the Blue Ridge that suggested prolonged maintenance of post-metamorphic temperatures in excess of those required for argon retention in higher grade parts of the Blue Ridge (Hadley, 1964; Dallmeyer, 1975). Alternative explanations include: (1) incomplete rejuvenation of detrital components at lowest metamorphic grades resulting in anomalously old apparent ages; and/or (2) polymetamorphism.

Geochronological resolution of detrial from authigenic very fine-grained white mica is difficult in very low-grade rocks. Dallmeyer and Takasu (1992) reported results of $^{40}Ar/^{39}Ar$ analyses of whole-rock slate/phyllite samples from a progressively metamorphosed sedimentary sequence in the Narragansett basin (Massachusetts-Rhode Island). They demonstrated that very fine-grained, pelitic slate/phyllite whole-rock systems (similar to those described herein from the western Blue Ridge) were completely rejuvenated during late Paleozoic metamorphism at grades above the middle anchizone (using illite crystallinity calibrations identical to those described herein). If appropriate for the western Blue Ridge, these controls suggest that except for sample 1 (diagenesis zone), intracrystalline argon systems within detrital mica grains in all other samples were likely completely rejuvenated during Paleozoic metamorphism(s). This implies that the intermediate-temperature $^{40}Ar/^{39}Ar$ ages recorded by samples 2–11 relate to the time of metamorphic rejuvenation and/or an associated growth of newly-formed white mica. The results combine to suggest two distinct tectonothermal episodes: one at 440 to 460 Ma (recorded in samples 2–5) and another at 340 to 380 Ma (recorded in samples 8–14). The latter thermal event apparently partially rejuvenated intracrystalline argon systems formed during the earlier event in samples 6 and 7. Muscovite separates (samples 12–14) from the staurolite and kyanite zones display slightly older ages (360–380 Ma) than whole-rock samples (340–350 Ma) as a result of higher temperatures of argon retention for muscovite. The muscovite results are therefore believed to date closely the younger metamorphic event (growth of prophyroblasts) in this area.

Intepretation of the geochronologic results as a record of polymetamorphic evolution is consistent with textural characteristics described earlier. Although the distribution of metamorphic isograds suggests a progressive Barrovian-type metamorphism (fig. 3), metamorphic textures from the study area and adjacent areas document a polymetamorphic history (Hadley and Goldsmith, 1963; King, 1964; Power and

Forrest, 1971; Mohr, 1973; Labotka and Shireman, 1991). Hadley and Goldsmith (1963) recognized that peak metamorphic conditions were preceded by and followed by deformation and lower grade metamorphism. The regional S_1 cleavage developed during an early kinematic phase of deformation. Porphyroblasts overgrew S_1 cleavage during a higher grade static phase of metamorphism. Porphyroblast growth was followed by S_2 crenulation cleavage that forms a penetrative schistosity at higher metamorphic grades. It is therefore likely that the 440 to 460 Ma whole-rock ages recorded within the chlorite metamorphic zone (where textural evidence for polymetamorphism is lacking) record the development of the regional S_1 cleavage in the western Blue Ridge. The younger 340 to 380 Ma ages recorded within biotite and higher metamorphic zones are interpreted to date the time of growth of porphyroblasts and thus the peak thermal conditions in this area. Together the $^{40}Ar/^{39}Ar$ results suggest that rocks of the western Blue Ridge experienced major periods of metamorphism at 440 to 460 Ma and at 360 to 380 Ma.

Thus, despite the apparently simple metamorphic isograd pattern in the western Blue Ridge (fig. 2), the present geochronologic results and metamorphic textures suggest a polymetamorphic history for this area. Unlike areas such as the northern Appalachians, however, detailed metamorphic petrology has only locally been conducted in the western Blue Ridge (Nesbitt and Essene, 1982; Mohr and Newton, 1983; Eckert, Hatcher, and Mohr, 1989). Resolution of areas where peak Paleozoic metamorphic conditions occurred in the early Paleozoic from areas that experienced peak conditions during a later period is not yet possible and awaits detailed metamorphic petrology.

TECTONIC IMPLICATIONS

Interpreting the geologic significance of the $^{40}Ar/^{39}Ar$ results depends upon calibration of the Paleozoic time-scale (Palmer, 1983; Harland and others, 1989). Snelling (1985) and Kunk and others (1985) suggested that the Ordovician-Silurian boundary (base of the Llandovery) is 435 to 440 Ma. This together with a 420 Ma calibration of the Ludlow (Wyborn and others, 1982) is used for interpretation of the $^{40}Ar/^{39}Ar$ results from the western Blue Ridge. McKerrow, Lambert, and Chamberlain (1980) proposed that the middle Devonian is bracketed by 396 and 382 Ma and that the Silurian-Devonian boundary is 412 Ma. Gale, Beckinsale, and Wadge (1980) discussed this calibration and suggested a compromise that defined the base of the Devonian at 400 Ma and the Middle Devonian to be bracketed by 387 and 374 Ma. This proposition is consistent with more recent time-scale calibrations (Palmer, 1983; Harland and others, 1989) and is used here.

Samples 2 to 5 were metamorphosed to conditions of the middle/upper anchizone. These should be near or below closure temperature of Ar in muscovite (Dallmeyer and Takasu, 1992). The 400 to 460 Ma ages from these whole-rock samples therefore closely date metamorphism. The ages are in general agreement with, although slightly younger than, the filling of the presently adjacent Sevier foreland by Blountian synorogenic sediments. Initial downwarping of the Sevier basin (reflected by deposition of the Whitesburg Formation; Shanmugan and Lash, 1982) occurred in the early Llanvirnian (475–480 Ma; Drak and others, 1989) and probably reflected eastward loading of the crust by thrust sheets. The major premetamorphic faults in this area (Dunn Creek, Greenbrier, and Hayesville) may have propagated at this time. Downwarping was followed by deposition of pelagic (Blockhouse Formation) and distal turbidite (Sevier Formation) sediments (Shanmugan and Lash, 1982) during the early Llanvirnian to middle Landeilian (475–465 Ma; Drake and others, 1989). The ages reported here, however, correspond most closely to deposition of the overlying molasse deposits (Bays Formation), which were deposited during the middle Llandeilian to early Caradocian (465–455 Ma; Drake and others, 1989). A similar correspondence between radiometric uplift ages and molasse sedimentation has been recognized in other orogenic belts (Trümpy, 1973; Eisbacher and Gabrielse, 1975).

A Late Devonian to earliest Mississippian paleontologic age assignment for regionally metmorphosed Walden Creek Group rocks of the western Blue Ridge by Unrug, Unrug, and Palmes (1991) is inconsistent with results from the present study that indicate a Middle to Late Ordovician metamorphism. It is also inconsistent with most previous geochronologic results in the western Blue Ridge, as well as previous paleontologic results and geologic mapping. The present $^{40}Ar/^{39}Ar$ results require that the Walden Creek Group be no younger than Late Ordovician, although stratigraphic arguments require an earlier (pre-Chilhowee) age. A Middle to Late Ordovician metamorphic age is also inconsistent with proposed stratigraphic correlations between rocks of the Murphy belt and Early Devonian strata of the Talladega belt (Tull and Guthrie, 1985; Thompson and Tull, 1991). The present results require that all Murphy belt units are no younger than Late Ordovician.

The 340 to 380 Ma whole-rock and muscovite ages are interpreted to indicate a Middle-Late Devonian tectonothermal event. The nature and importance of middle Paleozoic tectonism is poorly understood in the southern Appalachians (Ferrill and Thomas, 1988; Osberg and others, 1989; Tull and Telle, 1989). In the northern and central Appalachians, a thick Devonian clastic sequence in the foreland is temporally associated with metamorphism in the hinterland (Osberg and others, 1989). In the southern Appalachians, however, only a thin Lower to Middle Devonian clastic sequence (Frog Mountain Formation) is locally preserved in the foreland (Ferrill and Thomas, 1988). Within the Tennessee foreland, most of the Devonian underlying the sub-Chattanooga shale unconformity is absent (Colton, 1970). Tull and others (1988), however, provided fossil evidence to link the thick Cambrian to Devonian predominately clastic sequence of the Talladega belt with the Appalachian foreland in Alabama. This suggests that the southernmost Appalachian foreland may have once also contained a similar cover sequence. The stratigraphic record in the southern Appalachians therefore may not be inconsistent with a Middle to Late Devonian tectonothermal event.

ACKNOWLEDGMENTS

This research was, in part, supported by a grant (R.D.D.) from the tectonics program of the U.S. National Science Foundation (EAR-9117491). William Dunne, Robert Hatcher, John Rodgers, Nicholas Woodward, and anonymous reviews are thanked for careful reviews of the manuscript.

APPENDIX I

Sample	Latitude	Longitude
1	35°48'27"	83°27'33"
2	35°48'04"	83°25'48"
3	35°47'35"	83°24'24"
4	35°47'15"	83°24'17"
5	35°45'47"	83°26'46"
6	35°45'27"	83°24'19"
7	35°43'38"	83°29'07"
8	35°43'10"	83°23'24"
9	35°43'03"	83°23'13"
10	35°42'25"	83°19'01"
11	35°39'03"	83°26'34"
12	35°34'40"	83°20'45"
13	35°32'38"	83°18'09"
14	35°31'23"	83°18'24"

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