A FIELD AND GEOCHRONOLOGIC STUDY OF MANTLED GNEISS DOMES IN CENTRAL NEW ENGLAND

Thesis by

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Granite has given rise, perhaps, to more discussion concerning its origin than any other formation. We generally see it constituting the fundamental rock, and however formed, we know it is the deepest layer in the crust of this globe to which man has penetrated. The limit of man's knowledge in any subject possesses a high interest, which is perhaps increased by its close neighborhood to the realms of imagination.

Charles Darwin

December 30, 1834

<u>Voyage of the Beagle</u>

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ABSTRACT

New information on the origin of mantled gneiss domes comes from a study of these structures in Central New England. The domes have cores of massive granite and gneiss encircled by concordant mantles of well-stratified metamorphic rocks, and appear to originate through intense metamorphism of rock sequences in which massive, chiefly quartzo-feldspathic rocks are overlain by less competent strata. Contrary to previous hypotheses, the new work indicates that neither unconformable separation of the core and mantle nor re-mobilization or anatexis of the core rocks are essential elements in the formation of mantled gneiss domes. Two contrasting types of gneiss domes have been identified in central New England.

Examples of the first type are domes of the Chester Dome group in southeastern Vermont. Formation of these domes involved kyanite-staurolite grade metamorphism of Precambrian gneissic basement overlain unconformably by Paleozoic strata. The angular relationships at the unconformity have been obscured by differential movement of the core-rocks relative to the mantling strata. The Precambrian rocks in the cores of the Chester Dome and the nearby Green Mountain Anticlinorium have been badly disturbed by Paleozoic metamorphism, but generally yield Precambrian zircon and Rb-Sr whole-rock ages.

The second type of gneiss dome is exemplified by the Mascoma and Lebanon (Oliverian) Domes exposed about thirty miles to the east

in central New Hampshire. No Precambrian rocks have been identified in the cores of these domes. Fieldwork indicates that the core of the Mascoma Dome can be subdivided into two major units: (1) massive gneiss of intermediate igneous composition lying stratigraphically beneath the Ordovician Ammonoosuc Volcanics, and (2) a sub-central pluton of granite and quartz monzonite which crosscuts the massive gneiss and probably the Ammonoosuc Volcanics, but which lies unconformably beneath the Late Lower Silurian Clough Formation. Within limits imposed by analytical uncertainty and the metamorphic disturbance of the rocks, a common age of 440 ± 40 million years (initial $Sr^{87}/Sr^{86} = 0.706 \pm 0.002$) is determined for whole-rock samples of the granitic sub-cores of the Lebanon and Mascoma Domes, and for whole-rock samples of the Ammonoosuc Volcanics. Zircon separates from both the gneissic and granitic units within the core of the Mascoma Dome yield Pb^{207}/Pb^{206} ages of 450 ± 25 million years. The data indicate that these domes formed in the following stages: (1) Ordovician volcanism followed by intrusion of granitic rocks, (2) uplift and local unroofing followed by deposition of Lower Silurian through Lower Devonian strata, and (3) garnet- to staurolite-grade post-Lower Devonian metamorphism and deformation. Most of the crosscutting relationships were established by Ordovician plutonic activity and not by post-Lower Devonian plutonic activity or anatexis. The core-rocks of these domes appear to be the result of volcanic and intrusive activity towards the end of the Ordovician, and not the result of in-place remobilization or anatexis of Precambrian basement subsequent to deposition of the mantling strata. The other Oliverian Domes, particularly those in New Hampshire, resemble the Mascoma Dome, and probably originated in much the same manner.

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INTRODUCTION

The Oliverian Problem

Advances in geochronology have made it feasible to apply isotopic dating methods to the study of rocks which have been affected by multiple periods of thermal disturbance. One of the most interesting problems which can be studied with such techniques is the problem of the origin of mantled gneiss domes. The prevailing theory of their origin (Eskola, 1949) is that the domes have cores of older rocks that have been remobilized or even partly melted during intense later deformation. Field and geochronologic study by previous authors has shown that many mantled gneiss domes have cores of older rocks strongly affected by later disturbances. Other domes have yielded no evidence for the presence of older core-rocks, but it has been uncertain whether such rocks are absent or have had their geologic and isotopic "clocks" completely reset by later disturbance. The Oliverian Domes of New England were cited by Eskola (1949) as typical mantled gneiss domes and are well suited for further study of the mantled gneiss dome problem.

The Oliverian Domes are exposed in a belt which coincides approximately with the crest of the Bronson Hill Anticlinorium a few miles east of the Connecticut River. About twenty separate domes lie en-echelon along this belt from Long Island Sound to northeastern New Hampshire, a distance of about 250 miles. Each individual dome consists of a core of granite and gneiss encircled concordantly by a mantle of well-stratified metamorphic rocks. The core rocks of these domes are igneous in composition and appearance and locally have been reported to cut across the mantling strata. Following Billings (1937, p.502) most previous workers have interpreted all of the core rocks as granitic rocks intruded during the Devonian (Acadian) orogeny. Although a variety of intrusive mechanisms have been proposed, it has always been difficult to explain why the Oliverian core rocks are almost always in contact with a single unit of Ordovician metavolcanic rocks (Ammonoosuc Volcanics) along the entire 250 mile belt of domes. This is more the behavior to be expected of an older rock unit lying stratigraphically beneath the Ammonoosuc Volcanics and cropping out in a series of domical anticlines. This view was taken by Eskola (1949) in his classical paper on the origin of mantled gneiss domes, with the further suggestion that the rare crosscutting features were the result of remobilization of the older rocks during intense later metamorphism. Following the recognition of Precambrian rocks in the core of the nearby Chester Dome, a number of geologists (eg. Robinson, 1963) have suggested that some of the Oliverian core rocks may be Precambrian also. The difficulties of such an interpretation have been stressed by Billings (1956, p. 123-125) who maintained that evidence for the presence of remobilized rocks is lacking.

The core-rocks show a mixture of features characteristic of intrusive rocks with other features characteristic of stratified rocks. Attempts to explain all of the core-rocks as intrusive have failed to account for the stratigraphic characteristics, and conversely. The present author will attempt to show that the core-rocks can be subdivided in such a way that the intrusive and stratigraphic characteristics can be separated and attributed to discrete rock units. The basis for this separation has been established by mapping in the Mascoma area east of Hanover, New Hampshire. In the Mascoma Dome most of the core-rocks comprise a unit, the Holts Ledge Gneiss (new stratigraphic name), which lies stratigraphically beneath the Ammonoosuc Volcanics. Other core-rocks, the Mascoma Group, are probably intrusive. They crosscut the Holts Ledge Gneiss and possibly the Ammonoosuc Volcanics as well, but are older than the Clough Formation and higher mantle-units. The other Oliverian Domes, particularly those in New Hampshire, contain similar rocks, and probably originated in much the same manner as the Mascoma Dome. Age determinations further show that none of the core-rocks are as old as Precambrian and indicate that remobilization or granitization during the Acadian orogeny (the major period of deformation and metamorphism recognized in the area) did not play a significant role in the formation of the domes studied.

The author suggests the term "Oliverian" should be discontinued as the name of a rock unit. Billings (1937, p.501) applied the name "Oliverian Plutonic Series" to all light-colored rocks of igneous composition and appearance occurring beneath the Ammonoosuc Volcanics in the cores of the Oliverian Domes. Further usage of this term is inappropriate following the present author's conclusion that many of the core rocks are not intrusive, and further work is needed to determine to what extent the rocks of the dome belt constitute a related rock series. It seems proper to retain the name to refer to the "Oliverian Domes" or to the "Oliverian core-rocks".

Purpose and Scope

The chief purpose of this research has been to apply geologic and isotopic dating techniques to a study of the primary age relationships among rocks associated with the Oliverian mantled gneiss domes. Following a brief reconnaissance, the large Mascoma Dome east of Hanover, New Hampshire (Figures 1 and 2), was selected for detailed study. The Mascoma Dome is better exposed than most of the domes further north and contains most of the characteristic Oliverian rock types. Rocks in the area studied are less intensely metamorphosed and deformed than in most appropriate areas further south. Because the area contains one of two known localities where the Oliverian core-rocks are in contact with units stratigraphically higher than the Ammonoosuc Volcanics it seemed to be a suitable place to study possible crosscutting relationships of the core-rocks. Rb-Sr and U-Pb age determinations have been made on mineral and whole-rock samples from the core of the Mascoma Dome, and selected samples from the core of the Lebanon Dome, several Highlandcroft plutons, and from the Ammonoosuc Volcanics have been analyzed by the Rb-Sr method alone.

The author spent four summer months in 1964 and 1965 mapping contacts in the northwestern quadrant of the Mascoma Dome at 1:10,000 scale. Additional field-time was spent collecting samples, studying other domes, and examining the field evidence for critical geological relationships reported by previous workers in the area. The units mapped by previous geologists have been accepted with only slight modification, but the findings require considerable revision

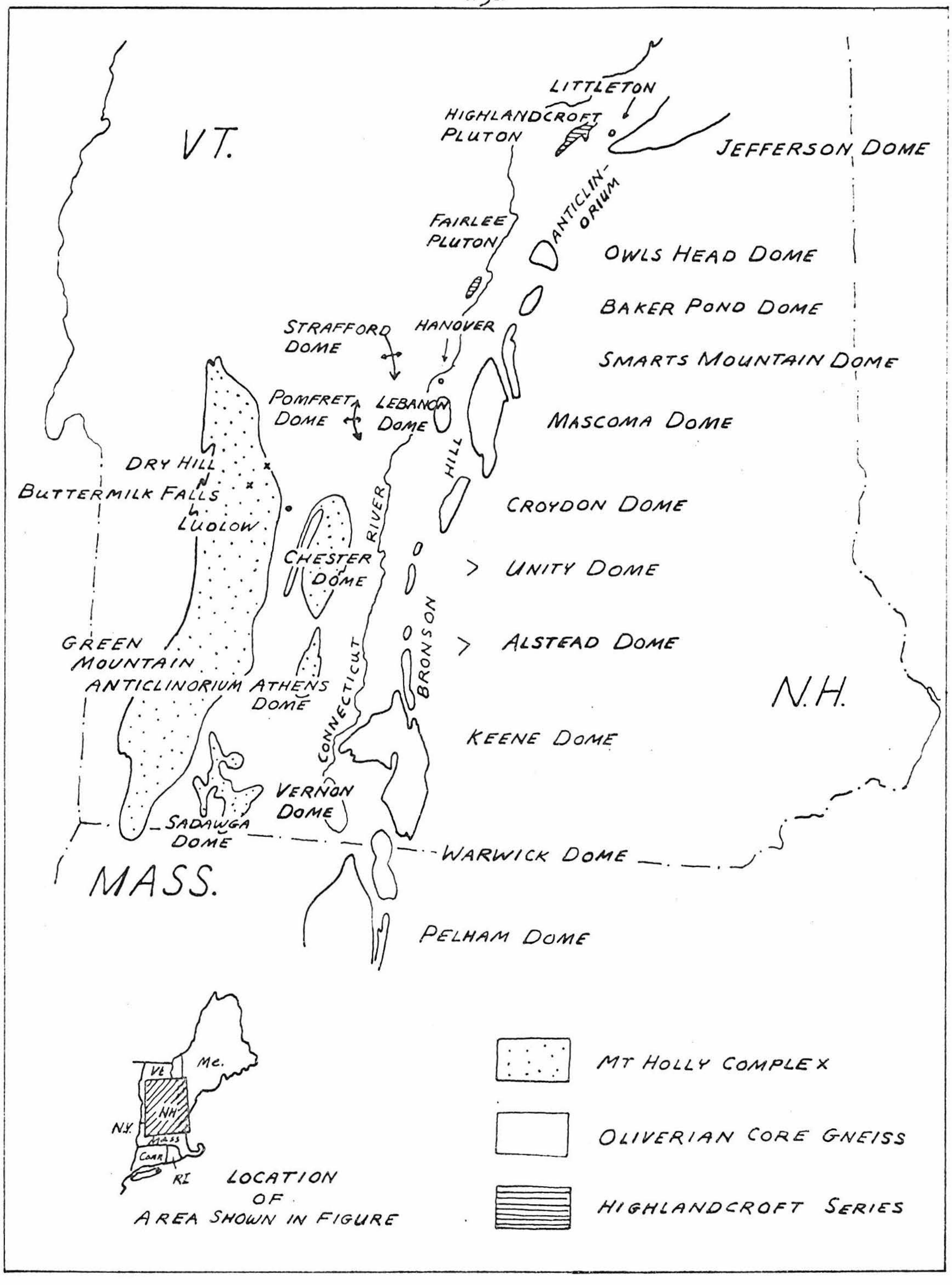


Figure 1. Localities mentioned in text.

of the age relationships among some of these units, and some units previously accepted as intrusive are reinterpreted as stratified formations. Previous workers have published thin-section modes, and in some cases chemical analyses, of samples from these rock units. References to these data are cited in Table 1 and Figure 12.

Contacts in the northern part of the Mascoma Dome were mapped primarily on enlarged air photographs. Location on the densely wooded hillsides was aided by use of traverse lines blazed with orange flagging tape. The data were transferred to a planimetric base (Figures 3 and 4) for publication. The topography on the Mascoma 15-minute quadrangle published in 1929 is not sufficiently accurate at the scale of the airphotos to justify its transfer to the detail maps, but the reader should bear in mind that the relative positions of the contacts of the more shallow-dipping strata are strongly influenced by the topography. This effect rather than changes in the thickness of the strata accounts for most of the variations in the width of the mapped units.

ROCK UNITS IN THE MASCOMA AREA

STRATIFIED ROCKS

unit	age	lithology	thickness (feet)	references
LITTLETON FORMATION	Lower Devonian	(1) quartz-mica schist, garnet-mica schist, and staurolite schist	(1) 3000 +	Billings (1937) p. 475-497; Chapman (1939) p. 135-142; Hadley (1942) p. 127-134; Robinson (1963)
FITCH FORMATION	Upper Silurian	(1) arenaceous marble, calcareous quartzite, quartz conglomerate, calcareous biotite schist, mica schist, and lime silicate granulite	(1) 400 to 600	
CLOUGH FORMATION	Lower Silurian	bedded quartzite, quartz conglomerate, quartz mica schist, quartz-mica-garnet-(chlorite-) schist	(1) 400	
(unconformity)				
AMMONOOSUC VOLCANICS	Ordovician	epidote-amphibolite, plagioclase-quartz-biotite-(hornblende-) gneiss, felsite, chlorite schist	400 +	
HOLTS LEDGE GNEISS	Ordovician (?)	plagioclase-quartz-biotite-(epidote-hornblende-) gneiss, felsite	2100 +	

UNSTRATIFIED ROCKS

unit	age	lithology	occurrence	references
MASCOMA GROUP	Early Silurian or Late Ordovician			Billings (1937) p. 501-502; Chapman (1939) p 142-146, 166-170; (1947) p. 599-613; Hadley (1942) p. 136-143 p. 159-164.
		granite and quartz monzonite	large, homogeneous pluton (Mascoma pluton)	
		quartz diorite, fine-grained granite, pegmatite	small, discordant intrusive bodies, and sills	

NOTE
(1) from Hadley (1942)

Table 1. Rock units in the Mascoma area, N.H.

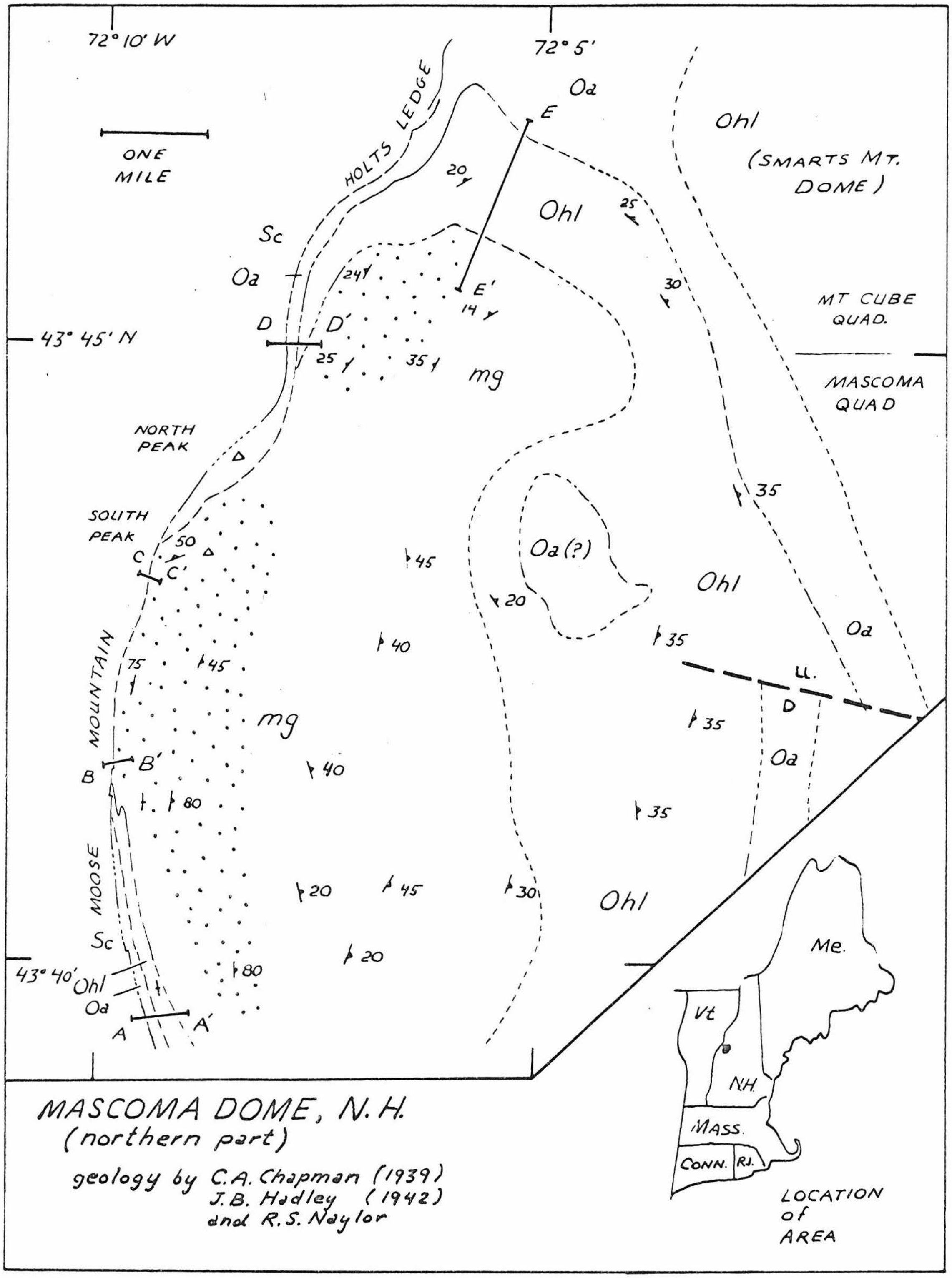


Figure 2. Geology of the northern part of the Mascoma Dome, N.H.

See Figure 4 for Explanation.

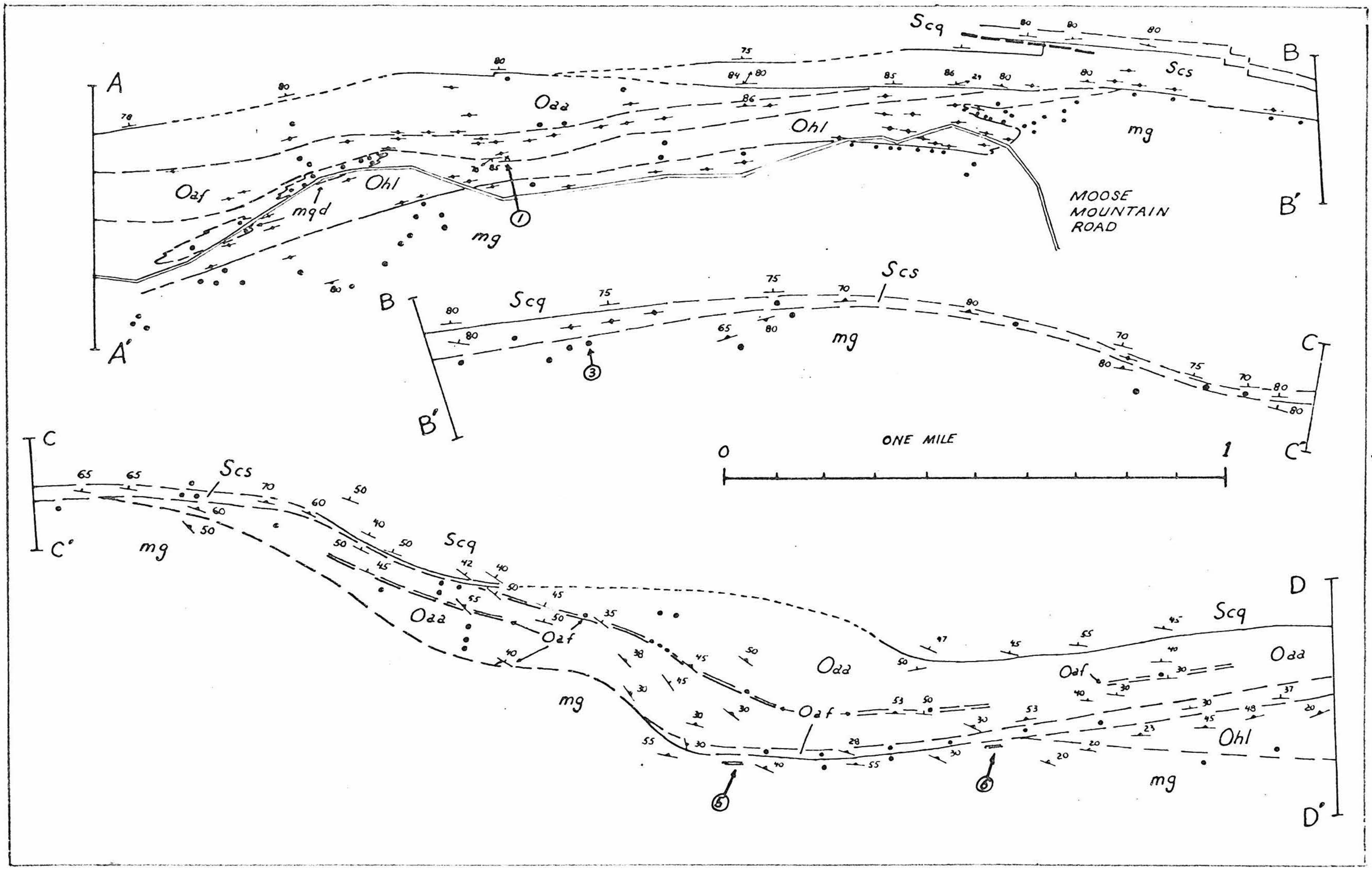


Figure 3. Geologic map of Moose Mt. (Explanation on Figure 4)

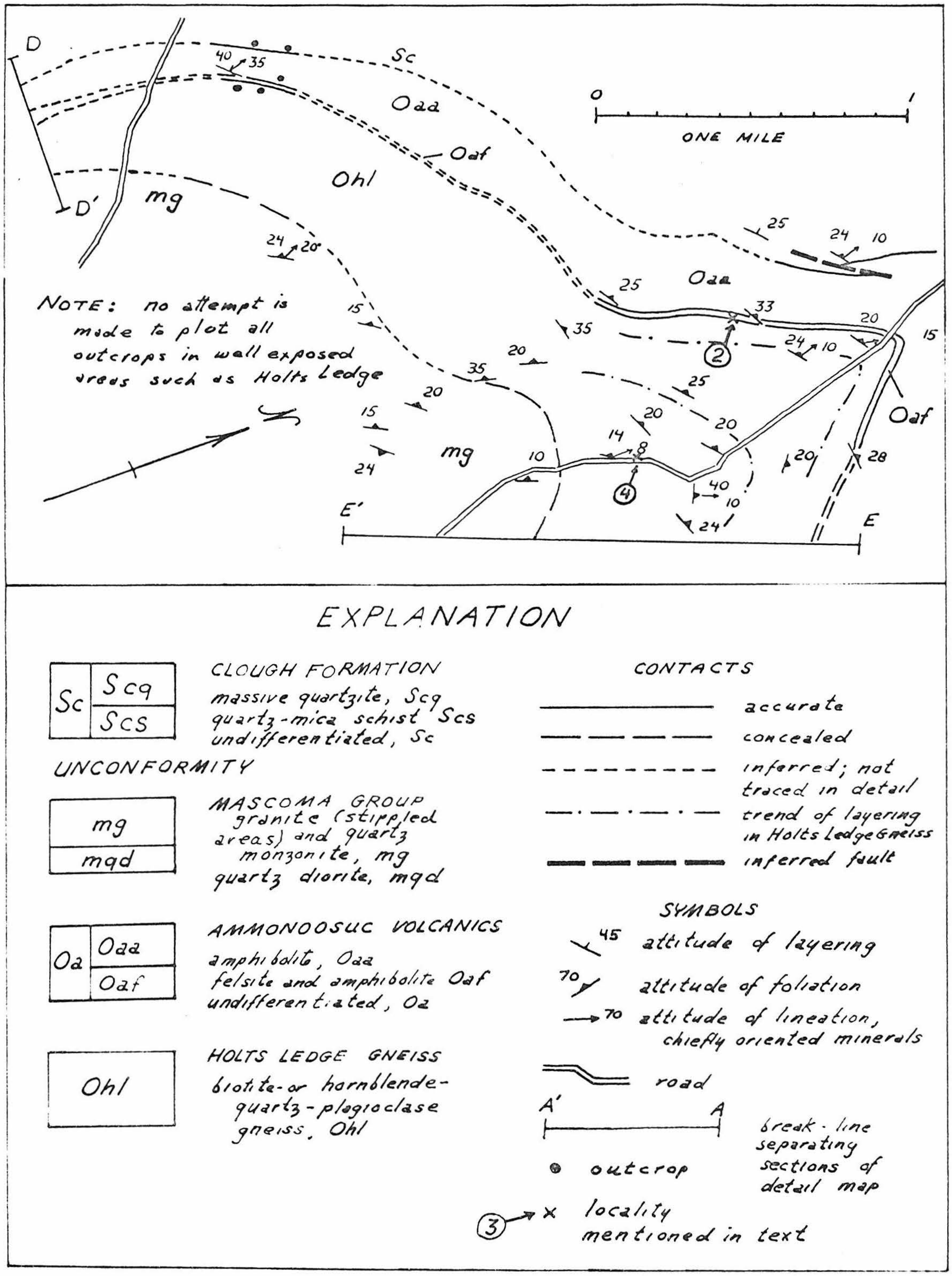


Figure 4. Geologic map of Holts Ledge area. (see Figure 1 for location)

GEOLOGY OF THE MASCOMA DOME

The Mascoma Dome (Figure 2), like the other Oliverian Domes, consists of a core of massive, light-colored gneiss and granite encircled concordantly by a mantle of well-stratified metamorphic rocks. The contrast between the core- and mantle-rocks is striking. The core-rocks are competent and behaved as a buttress during deformation. The author believes their structural behavior is chiefly responsible for the structure of the domes, and that the less competent mantle-rocks played an essentially passive role. Despite middle- and high-grade regional metamorphism, stratification is clear-cut in the mantle-rocks, but is weak or absent in the core-rocks. The stratigraphy of the mantle-rocks has been traced most of the length of the Oliverian Dome Belt, whereas the core-rocks are not connected at the surface and it has not been possible to trace stratigraphic elements from the core of one dome to the next. Datable fossils occur in some of the mantle-rocks and most geologists agree that the mantle-rocks originated through metamorphism of common sedimentary and volcanic stratified rocks. There has been no corresponding agreement regarding the origin of the core-rocks, for which previous geologists have suggested various hypotheses including igneous intrusion, metamorphism of older basement rocks, and remobilization or granitization. There has not even been agreement as to whether the core rocks are older or younger than the mantling strata. Most previous geologists have interpreted the lower contact of the Ammonoosuc Volcanics as a significant break, but have not agreed on its nature. This surface is considered to be the contact between the core and the mantle of the gneiss dome, and the Ammonoosuc Volcanics are considered to be the lowest unit of the mantle.

The core of the Mascoma Dome is a doubly-plunging asymmetric anticline. The west flank is steep and locally overturned, but the outward dip is more shallow on the other flanks. The lineation shown on Figures 3 and 4 is produced by a nearly down-dip orientation of the elongate mineral grains and aggregates. Foliation is generally parallel to stratification, and is most strongly developed in the stratified rocks. A weaker, but parallel, foliation is present in the unstratified rocks, and is somewhat better developed near the margins of the core than in the interior. Stretched quartz aggregates in the Holts Ledge Gneiss and the Ammonoosuc Volcanics and stretched quartzite cobbles in the Clough Formation suggest these units were thinned during the folding and metamorphism. Minor folds are rare in the core and lower parts of the mantle, but tight minor folds are found in the Clough and younger formations (J. B. Thompson, Jr., oral communication). Several breaks in the Clough Formation are shown in Figures 3 and 4. The author believes these breaks were caused by faulting, but they were not studied in detail. Around the Chester Dome group in southeastern Vermont (See Appendix 4) reverse-sense drag-folds, rotated garnets, and other structural features suggest that the core gneiss rose differentially with respect to the mantling strata (Thompson, 1950; Skehan, 1961; Rosenfeld, in preparation). No features were found to suggest this occurred in the Mascoma Dome. It appears to be a conventional anticline.

Mantle-Units

The mantle includes all units flanking the Mascoma Dome which lie stratigraphically above the base of the Ammonoosuc Volcanics. Geologists generally agree that the mantle-rocks originated through metamorphism (garnet- to sillimanite-grade in the Mascoma area) of sedimentary and volcanic stratified rocks. The units (see Table 1) have been described in considerable detail by previous workers, and the present treatment is limited to discussion of particular features of the lower units which are critical to the problem of the origin of the domes.

Clough Formation

The Clough Formation was defined by Billings (1937, p. 481-483) and has been described in the Mascoma area by Chapman (1939, p. 137-139) and Hadley (1942, p. 129-130). In the Mascoma area Boucot and Thompson (1963) have described well-preserved, highly-metamorphosed fossils which indicate the Clough Formation is of late Lower Silurian (C_4 to C_6) age.

The basal part of the Clough Formation has been mapped in this study to provide stratigraphic control above the Ammonoosuc Volcanics. The most characteristic rock is metasedimentary quartzite occurring in massive beds three to fifteen feet in thickness separated by thin partings or thicker interbeds of muscovite schist. Some of the beds are conglomeratic and contain abundant stretched quartzite cobbles. The basal unit of the Clough Formation is locally quartz-muscovite-garnet schist. Similar schist is interbedded with the quartzite higher in the section. At a few places near the contact with the Mascoma Group thin basal layers of the schist contain up to 50 percent dark tourmaline, and above the contact with the Ammonoosuc Volcanics the basal schist is commonly rich in chlorite.

Unconformity at the Base of the Clough Formation

The lower contact of the Clough Formation is a slight angular unconformity. The unit rests in apparent depositional contact on several of the underlying units (the Ammonoosuc Volcanics, the Mascoma Group, and probably the Holts Ledge Gneiss; see Figures 3 and 4). The lower contact of the Clough Formation truncates a crosscutting contact between rocks of the Mascoma Group and the Holts Ledge Gneiss (Figure 3), and there is no evidence that any of the Clough Formation is cut by the plutonic rocks. On Moose Mountain (Figure 3) the contact between the Ammonoosuc Volcanics and the Holts Ledge Gneiss appears to strike into the unconformity, although this relationship is not seen precisely at the contact. A number of peculiar rocks just beneath the Clough Formation suggest that the latter may have been deposited on a weathered zone developed in the older rocks. Locally (eg. Locality 3, Figure 3) the granite within a few feet of the contact with the Clough Formation contains chalky feldspars and about 20 percent muscovite. A possible explanation of this occurrence would be metamorphism of weathered granite which had been enriched in alumina by preferential removal of silica. Micaceous dark schists occur sporadically in the Ammonoosuc Volcanics in the first few feet beneath the Clough Formation.

Either an unconformity or a disconformity near the base of the Silurian is common throughout New England (Albee, 1961, p. C53; Billings, 1937, p. 483; Boucot, Field, Fletcher, Forbes, Naylor, and Pavlides, 1964, p. 88-93; Doll, Cady, Thompson, and Billings, 1961; Eaton and Rosenfeld, 1960, p. 171). At several localities besides the Mascoma area basal Silurian clastic units rest directly on underlying plutons or contain locally-derived clasts of the plutonic rocks (Albee and Boudette, in press; Pavlides, Mencher, Naylor, and Boucot, 1964, p. C35).

Ammonoosuc Volcanics

The Ammonoosuc Volcanics were defined by Billings (1937, p. 475-480) who assigned to them an Ordovician age. His usage has been followed substantially by most subsequent workers. The formation in the Mascoma area has been described at some length by Chapman (1939, p. 135-137) and Hadley (1942, p. 127-128). In the area mapped, the formation consists chiefly of layered epidote-bearing amphibolite which probably originated from metamorphism of mafic pyroclastic and extrusive volcanic rocks (Billings, 1937, p. 476). Some layers contain abundant lenses of epidote which Robinson (1963) suggested originated as volcanic bombs. Gray and bluish-gray quartz phenocrysts are common in many layers.

Sporadic felsite layers provide a measure of stratigraphic control within the Ammonoosuc Volcanics. The felsite is a light-buff-colored, plagioclase-quartz rock, commonly bearing streaks of iron oxide which may represent stretched amygdules. Some felsite layers contain gray or bluish-gray quartz crystals which may be phenocrysts.

The felsite layers range in thickness from a few inches to several feet and are interlayered with typical Ammonoosuc amphibolite. Similar felsite layers occur in nearby, less-metamorphosed exposures of the Ammonoosuc Volcanics where Billings (1937, p. 476-477) has argued convincingly that they originated chiefly as tuff and tuff-breccia. The parallelism of the felsite layers with the compositional layering in the amphibolite suggests that they are primary stratigraphic layers. The felsite is composed chiefly of millimeter-sized grains of quartz and oligoclase with a granoblastic or weakly gneissic texture. This texture is the likely result of metamorphism of a fine-grained volcanic rock. Mortar texture, the laminar texture characteristic of mylonites, or other features which might suggest that the felsite was produced by granulation of originally coarser rocks were not observed.

A persistent zone of felsite averaging about twenty feet in thickness has been traced about six miles along strike near the base of the Ammonoosuc Volcanics (Figures 3 and 4). The presence of such a marker horizon is important in the discussion of the concordance of the contact between the core- and mantle-rocks. Similar felsite horizons occur higher in the formation, but it has not proved possible to trace them for long distances along-strike.

The Ammonoosuc Volcanics rest on two underlying units, the Holts Ledge Gneiss and the Mascoma Group. On North Peak (Locality 5, Figure 3) typical quartz monzonite of Mascoma Group is exposed within five feet of typical Ammonoosuc amphibolite. Elsewhere the contact is not sufficiently exposed to establish whether the Ammonoosuc Volcanics are crosscut by the Mascoma Group. The contact of the Ammonoosuc Volcanics with the Holts Ledge Gneiss appears concordant. It is described later as the upper contact of the Holts Ledge Gneiss.

Figure 2 shows an area underlain by dark amphibolite labled Oa (?). This was interpreted by Chapman as Ammonoosuc Volcanics, but it is possible that the rocks are amphibolite layers within the Holts Ledge Gneiss.

Core-units

The origin of the core-units has long been problematical. Following Billings (1937, p. 501) most geologists who have mapped the Oliverian Domes have interpreted all of the core-rocks as intrusive, but this explanation has not proved entirely satisfactory. In addition to those features which suggest an igneous origin, the core-rocks show other features characteristic of stratified rocks. The present author's mapping demonstrates that it is possible to subdivide the core-rocks of the Mascoma Dome into two major units. Both of these units are igneous in composition, but only one of the units, the Mascoma Group, unequivocally crosscuts other units. The other unit, which is assigned the new stratigraphic name, Holts Ledge Gneiss, is weakly stratified and has nowhere been observed to crosscut other units. In the discussion which follows, the author suggests that the Holts Ledge Gneiss may have originated through metamorphism of a sequence of intermediate volcanic rocks, whereas the rocks of the Mascoma Group are probably intrusive.

Holts Ledge Gneiss

In the northern part of the Mascoma Dome the predominantly mafic Ammonoosuc Volcanics give way down-section to a great thickness of weakly-stratified gneiss of intermediate igneous composition which constitutes much of the core-rock of the dome. The transition is not abrupt as there is a zone about a hundred feet thick in which sporadic layers of Ammonoosuc-type amphibolite alternate with layers of the light-colored gneiss, and some of the layers of the gneiss are lithologically similar to some of the felsite layers in the Ammonoosuc Volcanics. For the following reasons the author believes the intermediate gneiss should be recognized as a distinct map-unit. (1) The massive character of the gneiss and its structural position require that it be considered part of the core-rock of the Mascoma Dome, while it is logical to assign the better-stratified Ammonoosuc Volcanics to the mantle. (2) Previous authors (Chapman, 1939; Hadley, 1942) have argued that the gneiss is intrusive, and it is therefore necessary to consider its origin separately from that of the Ammonoosuc Volcanics. (3) The gneiss is lithologically and texturally distinct from the granite and quartz monzonite which the author assigns to the Mascoma Group. Page (1940) included several thousand feet of rock similar to the gneiss in the Ammonoosuc Volcanics at the south end of the Owls Head Dome (Figure 1), but this practice has not been followed by subsequent workers and is inconsistent with the criteria discussed above.

The author assigns the new stratigraphic name, Holts Ledge Gneiss, to the gneiss unit. The type area of the Holts Ledge Gneiss is at Holts Ledge south of Lyme Center, New Hampshire (Mt. Cube quadrangle). In the type area the exposed thickness of the unit is about 2100 feet. The layers do not appear to have been structurally duplicated and may have been tectonically thinned. The unit is intruded by rocks of the Mascoma Group, and a lower stratigraphic contact was nowhere seen in the Mascoma area. The upper contact is discussed in the following paragraphs.

In the type area the author places the upper contact of the Holts Ledge Gneiss at the surface below which the massive light-colored gneiss constitutes more than fifty percent of the gross section. A considerable length of this contact is exposed high on the vertical cliffs at Holts Ledge. One exposure of the contact which is more easily reached is indicated as Locality 2 on Figure 4 (43° 46.79' N; 72° 6.15' W). The following description is given to aid in identifying the contact in the field. About ten feet above the contact at Locality 2 is a natural terrace in the hill-slope. The lower felsite zone near the base of the Ammonoosuc Volcanics is exposed in the ledges which rise above this terrace. Between the felsite and the contact is about 30 feet of fine-grained, dark amphibolite. Below the contact amphibolite occurs only as sporadic layers which constitute a minor part of the section.

In choosing the definition of the contact the author was guided by the criteria discussed previously for establishing the identity of the Holts Ledge Gneiss. The definition is consistent with the criteria established by most previous geologists for distinguishing the core rocks of the Oliverian Domes from the Ammonoosuc Volcanics. It assigns to the core most of the rocks which Chapman (1939) and Hadley (1942) believed are intrusive, and retains their usage of the Ammonoosuc Volcanics. One exception is on Moose Mountain (near Locality 1, Figure 3) where the author includes in the Holts Ledge Gneiss about one hundred feet of strata which Chapman (1939) assigned to the Ammonoosuc Volcanics. The author's definition is consistent with criteria established by Robinson (1963) for recognizing the base of the Ammonoosuc Volcanics in the Orange area, Massachusetts.

The contact between the Holts Ledge Gneiss and the Ammonoosuc Volcanics is parallel to the internal stratification of both units. Figures 3 and 4 show that the gneiss maintains a constant stratigraphic position relative to the felsite horizons near the base of the Ammonoosuc Volcanics over the six mile stretch of the contact which was studied in detail. Binocular observation of the contact on the high cliffs at Holts Ledge likewise indicates the concordance of the contact. Concordance on a finer scale could not be verified where the contact is directly exposed due to lack of lamination in the Ammonoosuc amphibolite immediately above the contact.

The Holts Ledge Gneiss is weakly stratified. The gneiss is made up of massive layers averaging three to ten feet in thickness which are separated by discrete planes of discontinuity. Faint lamination is visible in some of the layers, but strongly banded gneiss is absent. Adjacent layers commonly differ slightly in composition or texture. The most common rock is massive, medium grained, light-colored plagioclase-quartz-biotite-epidote gneiss of intermediate igneous composition. The biotite characteristically occurs in oval-shaped aggregates about a centimeter across, which impart a distinctive blotchy appearance to many exposed foliation surfaces. In the lower part of the unit the differences between layers are mostly minor variations in the relative abundance of aggregates of quartz crystals, the presence or absence of minor amounts of garnet, magnetite, or muscovite, and the degree to which the biotite is aggregated. Rare concordant amphibolite layers occur in the lower part of the unit.

The upper part of the Holts Ledge Gneiss is more heterogeneous. The typical light-colored, plagioclase-quartz-biotite-epidote gneiss with oval-shaped aggregates of biotite persits to near the top of the section, but in the upper few hundred feet it is interlayered with gneisses having more strongly variant lithology and texture. A common variant is gneiss in which the biotite is less abundant and more uniformly distributed throughout the rock, and in which some surfaces parallel to the foliation are rich in hornblende laths with preferred linear orientation. Some layers contain abundant disc-shaped lenses of darker plagioclase-biotite gneiss averaging about one foot in greatest dimension. One layer has the texture and composition of meta-rhyolite and several other felsite layers of undetermined composition are present. The layers vary considerably in grain size but the average is intermediate between the medium-grained gneisses below and the fine-grained Ammonoosuc Volcanics above. Near the top are several concordant layers of amphibolite similar to the Ammonoosuc amphibolites, but which are overlain by rock typical of the Holts Ledge Gneiss. At Holts Ledge one of these amphibolite layers about five feet thick can be traced laterally for about two hundred feet. The layer is precisely concordant, and is nowhere intruded by the more typical Holts Ledge Gneiss.

The average composition of the Holts Ledge Gneiss changes from quartz diorite high in the section to granodionite in the lower parts. The change is not abrupt as there are sporadic interlayers having granodionitic or even granitic composition even in the uppermost parts of the unit. The change in composition is not correlated with any change in the nature of the stratification or the general appearance of the rocks and subdivision based on this change in composition would be very difficult to map in the field.

Mascoma Group

As discussed previously it is possible to distinguish from the Holts Ledge Gneiss other core-rocks which are more massive, which lack even the faint stratification noted in the gneiss, which are lithologically distinct, and which occur in crosscutting bodies. These unstratified rocks consist chiefly of quartz monzonite and granite, together with minor amounts of quartz diorite, fine-grained granite, and pegmatite. The name, Mascoma Group, is retained for these unstratified rocks, but is restricted from usage of Chapman (1939) and Hadley (1942) to exclude the intermediate gneiss which the present author assigns to the Holts Ledge Gneiss.

Most of the rocks of the Mascoma Group occur in a discrete, oval-shaped pluton of quartz monzonite and granite which the author calls the Mascoma Pluton. Most of the unit is porphyritic quartz monzonite, containing microcline in irregular crystals up to one centimeter across which impart a characteristic appearance to the rock. Plagioclase (An_{25-30}) and most of the quartz occurs as an equigranular, medium-grained matrix. Some of the quartz occurs in crushed aggregates, and the biotite and epidote are commonly also in aggregates. Locally the rocks contain small magnetite octahedra or small red garnets. An unusual type of exsolution occurs in the microcline. Albite ($Ab_{97}An_{3}Or_{<1}$) is exsolved in small 0.3 mm. patches within nearby pure microcline ($Or_{100}Ab_{<1}An_{<1}$). (The feldspar compositions were determined by microprobe analysis.) The crosshatch microcline twinning is continuous through both phases.

The granitic rocks of the Mascoma Pluton are massive and homogeneous on an outcrop scale. Modification of the primary texture of the granite during deformation includes some crushing of the groundmass minerals and quartz aggregates and the development of a moderate foliation which is strongest near the outer margins of the core. Allowing for such modification, the rocks of the pluton bear a strong resemblance to the massive granitic rocks of batholiths which are believed to have crystallized at intermediate depths. They lack the strongly foliated, streaked, veined, "juicy", or migmatitic appearance characteristic of deeper-seated granitic rocks. Sporadic veins of aplite with sharp crosscutting contacts occur locally within the pluton. The granitic rocks contain dark, elongate, plagioclase-quartz-biotite inclusions oriented parallel to the foliation. These inclusions are locally conspicuous but nowhere constitute as much as one percent of the rock.

There is a continuous variation in the abundance of microcline in the rock ranging from about twenty to fifty percent. The rocks with the most microcline have the composition of granite and generally contain less biotite and epidote than the quartz monzonite which constitutes most of the pluton. The granite is concentrated near the western margin of the pluton and was mapped separately by Chapman (1939) and Hadley (1942; areas indicated by stippling in Figure 2). The gradational nature of the variation within the Mascoma Pluton contrasts with the sharpness of the outer contacts of the pluton.

A minor portion of the Mascoma group consists of quartz-diorite, fine-grained granite, and pegmatite. The quartz diorite contains plagioclase, quartz, and biotite or hornblende. It is similar in composition to rocks common in the Holts Ledge Gneiss and is easy to recognize only where it occurs in bodies with crosscutting contacts as on southern Moose Mountain (Figure 3). The fine-grained granites have the same mineralogy as the medium-grained granites. Rare pegmatite veins contain large crystals of sodic plagioclase, microcline, quartz, and biotite.

Contact Relationships of the Mascoma Group

In contrast to the gradational variations of composition within the Mascoma Pluton, the body is separated from the Holts Ledge Gneiss, the Ammonoosuc Volcanics, and the Clough Formation by contacts which are sharp and readily traceable in the field. At several localities typical rocks of the pluton are exposed within a few feet of the other units, although the actual contacts were nowhere seen.

The granitic rocks of the Mascoma Pluton crosscut the Holts Ledge Gneiss. At one locality on Moose Mountain (Figure 3) layering in the Holts Ledge Gneiss is truncated by the granitic rocks, and a projecting sill of fine-grained quartz monzonite extends a short distance between layers of the gneiss. Figure 4 shows that the northern contact of the pluton cuts across the trend of probable stratigraphic layering in the Holts Ledge Gneiss. The composition of the Holts Ledge Gneiss is not noticeably altered near the contact with the granitic rocks. Near Locality 1 (Figure 3) typical quartz monzonite is exposed about ten feet from amphibolite in the Holts Ledge Gneiss. The amphibolite is cut by veins of quartz diorite, but does not appear to be enriched in potassium minerals.

The relationship between the rocks of the Mascoma Pluton and the Ammonoosuc Volcanics is uncertain. At Locality 5 (Figure 3) typical quartz monzonite is exposed about five feet below Ammonoosuc amphibolite. At Localities 5 and 6 (Figure 3) concordant layers of gneiss measuring about 5 by 100 feet occur in the granite about 40 feet below its contact with the Ammonoosuc Volcanics. The layers may be slabs of Holts Ledge Gneiss or Ammonoosuc Volcanics included in the quartz monzonite, or the overlying quartz monzonites may be sills. Southwest of Locality 5 (Figure 3) there is only one exposure of the persistent felsite horizon near the base of the Ammonoosuc Volcanics. Sufficient exposures are lacking in this area to make it certain that the felsite has been cut out by the quartz monzonite and granite. Nowhere do the Ammonoosuc Volcanics appear to have been altered in composition or texture near the granite.

The contact between the Mascoma Pluton and the Clough Formation is an angular unconformity. Chapman (1939) indicated that part of the Clough Formation had been cut by a large bulge of the granitic rocks at South Peak. Detailed mapping on an airphoto base shows that his contact in this area was misplaced due to inaccuracies in the topography on the 1927 edition of the Mascoma 15-minute quadrangle. The present author has traced the basal layers of the Clough Formation along the contact and there are no beds cut by the granitic rocks. The type exposure of Chapman's (1939, p. 168) "feldspathized Clough" (Locality 7; Figure 3) strongly resembles crushed, pyritized granite found in shear zones elsewhere in the dome. The exposure lies well within the granite about 100 feet below the lowest beds of the Clough Formation, and is not definitely in place. Chapman's inference that this material indicates the Clough Formation was feldspathized by the granite does not appear to be supportable.

Quartz diorite assigned to the Mascoma Group occurs in small bodies which crosscut layering in the upper part of the Holts Ledge Gneiss (see Figure 3). The rocks contain abundant, dark, biotite-rich xenoliths, and the amphibolite layers adjacent to the bodies are cut by intricate networks of quartz diorite veins. The quartz diorite has nowhere been observed to crosscut amphibolite definitely assigned to the Ammonoosuc Volcanics or to crosscut the granitic rocks of the Mascoma Group. The pegmatites are not strongly foliated and may be younger than the major periods of disturbance, but they were not observed to crosscut the mantle-rocks.

GEOLOGY OF RELATED ROCK UNITS

The Lebanon Dome

Rocks similar to those in the Mascoma Dome occur in the core of the Lebanon Dome, whose center lies about 8 miles west of the center of the Mascoma Dome. Like the Mascoma Dome, the Lebanon Dome has a sub-core of medium-grained granitic rocks bordered by a larger mass of granodioritic gneiss (Lyons, 1955, p. 118). The granitic rocks are similar in hand-specimen and thin-section to those in the core of the Mascoma Dome. Microcline has the same unusual type of perthite that was noted in microcline from the Mascoma Group, and biotite is aggregated in a manner similar to that in the Mascoma Group. The granitic rocks of the Lebanon Dome are better foliated and better lineated than those in the Mascoma Dome and appear to be somewhat more strongly deformed. The Lebanon Border Gneiss which underlies most of the core of the dome ranges in composition from quartz diorite to granodiorite (Lyons, 1955, p. 118). In composition and general appearance it is similar to the Holts Ledge Gneiss. The Border Gneiss is rather poorly exposed and the author has not studied it in sufficient detail to establish whether it is stratified in the same manner as the Holts Ledge Gneiss. The Lebanon Dome is mantled by feldspathic gneiss containing thin lentils of chlorite and hornblende schist which Chapman (1939, pl. 1) assigned to the Post Pond Volcanics. According to J. B. Thompson, Jr. (oral communication, 1964) the Post Pond Volcanics have the same stratigraphic position as the Ammonoosuc Volcanics. The contact relationships of the Lebanon Dome were not studied in detail by the present author.

Most previous geologists have correlated the core-rocks of the Lebanon Dome with those of the other Oliverian Domes. The granitic core-rocks of the Lebanon Dome strongly resemble those of the Mascoma Group, and the author feels that the minor differences may be due to somewhat greater deformation in the Lebanon Dome. The stratified rocks in New England show great persistence parallel to strike but change rapidly perpendicular to strike. The author feels that the mantle rocks of the Lebanon and Mascoma Domes are similar in stratigraphic position and that most of the lithological differences are due to facies changes.

Highlandcroft Series

Several plutons of the Highlandcroft Series (Billings, 1937, p. 499-500) are exposed north and northwest of the Mascoma Dome. Lahee (1913, p. 231-234) and Billings (1937, p. 500) have argued that the Highlandcroft Pluton west of Littleton, New Hampshire (see Figure 1) is intrusive into the Ammonoosuc Volcanics but lies unconformably beneath the Fitch Formation. The Lower Silurian Clough Formation is absent beneath the Fitch Formation in the Littleton area. The stratigraphic position of the Highlandcroft Pluton is therefore similar to that of the Mascoma Pluton, although it has not been possible to demonstrate whether the latter crosscuts the Ammonoosuc Volcanics. The composition of the Highlandcroft Pluton ranges from diorite and quartz diorite to quartz monzonite. Microcline commonly occurs in anhedral crystals up to a centimeter across similar to the phenocrysts (?) of the Mascoma Group. Bluish or gray quartz commonly occurs in elongate aggregates . The Highlandcroft rocks contain up to seven percent chlorite whereas the rocks of the Mascoma Group generally contain none. This mineralogic difference may be due to differences in the later metamorphic history of the rocks. The Fairlee quartz monzonite (see Figure 1) was correlated with the Highlandcroft Series by Hadley (1942, p. 136) chiefly on the basis of lithology. Hadley (1942, p. 136) states that the Fairlee Pluton cuts the Albee Formation which underlies the Ammonoosuc Volcanics. The lithology and field relationships of these Highlandcroft rock-units suggest they may be roughly correlative with the granitic rocks in the cores of the Oliverian Domes and they were sampled for dating purposes.

GEOCHRONOLOGICAL DATA

Whole-Rock Rb-Sr Determinations

Rb-Sr whole-rock determinations have been made on samples of the core-rocks of several Oliverian Domes, the Ammonoosuc Volcanics, and several plutons of the Highlandcroft Series. For Rb-Sr whole-rock dating, suites of samples must be obtained which satisfy the following conditions: (1) all of the rocks in the suite must have formed during a single interval of time which is short compared to their age; (2) at the time of their formation all of the rocks must have incorporated common Sr of the same isotopic composition; (3) subsequent to the time of formation, radioactive decay must be the only process acting to change the concentration of isotopes of Rb and Sr in the samples (closed system assumption); and (4) there must be a sufficient variation in the Rb/Sr ratios of the samples to establish a spread of points when the data are plotted. If these conditions are satisfied the data for the rocks will plot as a linear array (an isochron) on the strontium evolution diagram (see discussion by Lanphere, Wasserburg, Albee, and Tilton, 1964, p. 280-285). The age of the suite corresponds to the slope of the array, and the isotopic composition of common Sr incorporated at the time of formation is given by the intercept of the array on the Sr^{87}/Sr^{86} axis. If there is not much variation in the Rb/Sr ratios of samples within a suite, the data points cluster together and it is difficult to determine accurately the slope and intercept of the line to be fitted through them. If the samples have low Rb/Sr ratios, the buildup of radiogenic

Sr with time is slight, and corrections in the isotopic composition of the initial Sr produce large uncertainties in the determined ages. The effect of such uncertainties is less for samples with high Rb/Sr ratios as the correction for common Sr is smaller.

Most of the rock units of interest in this study are only marginally suitable for whole-rock dating. The Rb/Sr ratios are relatively low and show little variation among the common rock-types within the units. As the rocks are also relatively young, the enrichment in radiogenic Sr is slight. Sr in typical samples is only 1 to 3 percent radiogenic and large errors are introduced by uncertainties in the correction for common Sr. To improve the range of Rb/Sr ratios it has been necessary to process samples of rock-types such as aplite and fine-grained granite which are only minor constituents of the units to which they are assigned. The geologic suitability of such samples must be examined in each case. In what sense do they "belong" to the units to which they are assigned? Are they cogenetic, and did they incorporate common Sr of the same composition as the other rocks? How did they react during disturbance?

The aplites studied appear to be integral parts of the granitic cores of the Lebanon and Mascoma Domes, although the degree of metamorphism makes it difficult to prove this assertion. The metamorphism would be particularly effective in effacing evidence such as chilled contacts which might provide the most sensative indications of the relative age of the aplite veins. The aplites are intimately associated with the granites and probably formed during the later stages of the consolidation of the granites. No evidence was found suggesting the aplites are significantly younger. They are foliated to the same degree as the granites and were nowhere observed to crosscut units younger than the units crosscut by the granites. The fine-grained granites are less intimately related to the medium-grained granitic rocks, but share the same foliation and have not been observed to crosscut younger rocks. It is probable that any errors introduced by basing the isochrons heavily on these minor phases tend in the direction of lowering the apparent age of the rock units. The aplites and fine-grained granites are probably among the younger rocks in each unit, and some of the fine-grained granites may have been emplaced significantly later than the enclosing rocks. Disturbance during metamorphism could lower the apparent age of such units if they tended to exchange their Sr with the much less radiogenic Sr of the units in which they are embedded. The small size of the aplite and fine-grained granite bodies enhances the efficiency of this type of possible exchange.

The Rb-Sr analytical data are presented in Table 2. Analytical techniques, and the precision and accuracy of the measurements are discussed in Appendix 1. Descriptions of the samples are given in Appendix 2. All Sr data have been normalized to $(Sr^{86}/Sr^{88}) = 0.1194$, and the decay constant, $\lambda\, Rb^{87} = 1.39 \times 10^{-11} year^{-1}$ has been used. The well-known uncertainty in this decay constant does not affect the intercomparison of the Rb-Sr data, but must be considered when the ages are compared with U-Pb ages or discussed in terms of the geologic time-scale. The errors assigned to the data are calculated from a

RB-SR DATA

SAMPLE	PHASE	DATE ANALYZED	Rb^{87} (10^{-8} moles/gram)	Sr^{86} (10^{-8} moles/gram)	$\frac{Rb^{87}}{Sr^{86}}$	$\frac{Sr^{87}}{Sr^{86}}$	$\frac{Sr^{87}_{rad.}}{Sr^{87}}$	T (10^{12} sec)
Green Mountain Anticlinorium								
GDH-2	ω_I	2-65	54.22	11.24	4.82	0.779[4]	.09.14	3327
GDH-3	ω_1	3-65	42.82	6.02	7.12	0.834[4]	0.161	4278
GDH-4	ω_I	1-65	61.57	6.17	9.98	0.827	0.144	2699
GBF-7	ω_I	3-65	67.02	2.10	31.93	1.059	0.011	1337
GBF-8	ω_I	2-65	99.38	1.83	54.18	1.261	0.439	2305
GBF-9	ω_I	4-65	104.66	2.56	40.81	1.105	0.359	2195
GBF-10	ω_I	1-65	70.49	1.28	55.05	1.262	0.439	2273
GMH-4	ω_{III}	12-63	16.00	19.94	0.80	0.725	0.023	4649
	m	1-64	90.14	5.25	17.18	0.825	0.142	1540
	b	2-64	145.3	1.26	115.3	1.210	0.415	987
GMH-3	m	11-63	76.50	2.70	28.29	0.888	0.203	1438
Chester Dome Core								
CSG-1	ω_I	2-64	4.69	65.40	0.07	0.706[4]		
	b	5-64	80.05	4.74	16.90	0.799	0.114	1222
CSG-2	ω_I	3-64	24.57	36.97	0.67	0.708[4]		
	b	3-64	87.75	1.52	57.89	0.968	0.268	1016
CSG-4	ω_I	5-64			0.02[2]	0.702[4]		
CSG-5	ω_{II}	4-64	30.34	35.85	0.85	0.709		
	b	4-64	70.30	1.55	45.25	0.886	0.201	890
CSG-6	ω_I	5-64			0.04[2]	0.705[4]		

SEE EXPLANATION AT END OF TABLE

Table 2. Rb-Sr data

RB-SR DATA, CONT.

SAMPLE	PHASE	DATE ANALYZED	Rb^{87}	Sr^{86}	$\frac{Rb^{87}}{Sr^{86}}$	$\frac{Sr^{87}}{Sr^{86}}$ (3)	$\frac{Sr^{87}_{rad.}}{Sr^{87}}$	T (1) (10^{13} sec)
			(10^{-8} moles/gram)					
Lebanon Dome Border gneiss								
LBR-1	w_I	1-65	48.31	35.68	1.35	0.716 (4)	0.011	1337
Granitic core								
LCV-1	w_I	1-65	50.80	20.25	2.51	0.722 (4)	0.019	1263
LCQ-1	w_I	1-65	55.50	17.61	3.15	0.722	0.020	1021
	w_{IV}	4-67	58.99	17.87	3.30	0.722	0.020	963
LCR-1	w_I	12-64	47.49	8.42	5.64	0.740 (4)	0.046	1330
LCW-1	w_{II} (5)	3-66	80.95	1.72	46.99	0.979	0.277	1306
	w_{III}	1-67	81.07	1.67	48.53	0.985	0.281	1293
	w_{IV}	1-67	80.51	1.66	48.23	0.985	0.282	1302
LCR-3	w_{IV}	12-64	71.10	4.22	16.86	0.810	0.125	1368
	w_{VI}	3-66	72.36	4.23	17.11	0.810	0.125	1343
LCR-5	w_{III}	1-66	60.56	15.04	4.03	0.732	0.033	1367
	w_{IV}	12-64	59.78	14.50	4.12	0.733	0.034	1361
	ap	3-65	0.38	19.28	0.02	0.721	0.018	
	p	12-64	33.53	9.92	3.38	0.729	0.029	1406
	k_I	2-65	125.9	19.08	6.60	0.738	0.040	1013
	k_{III}	4-66	124.9	19.34	6.46	0.738	0.040	1038
	m	12-64	196.3	2.30	85.3	1.091	0.351	1018
	b	1-65	240.3	1.56	154.2	1.250	0.434	796
Mascoma Dome Holts Ledge Gneiss								
MWL-1	w_{II}	6-65	15.59	8.83	1.76	0.713	0.007	629
Mascoma Group (granitic core)								
MMQ-11	w_I	2-66	49.41	38.11	1.30	0.712	0.006	735
MMS-2	w_I (5)	6-65	48.41	42.44	1.14	0.716	0.012	1665
	w_{II}	1-66	48.33	40.95	1.18	0.716	0.011	1496
MMQ-3	w_I	5-65	48.75	15.76	3.09	0.722	0.019	1025
MCP-1	w_I	7-65	26.12	3.87	6.74	0.740	0.043	1078

SEE EXPLANATION AT END OF TABLE

Table 2, cont.

RB-SR DATA, CONCLUDED

SAMPLE	PHASE	DATE ANALYZED	Rb^{87} (10^{-8} moles/gram)	Sr^{86} (10^{-8} moles/gram)	$\frac{Rb^{87}}{Sr^{86}}$	$\frac{Sr^{87}}{Sr^{86}}$ (3)	$\frac{Sr^{87}_{rad.}}{Sr^{87}}$	T (1) (10^{13}sec)
Moscoma Dome, cont. Moscoma group, cont.								
MJH-1	w_I	4-65	20.54	6.23	3.30	0.725	0.023	1132
	m	2-66	102.3	2.07	49.45	0.904	0.217	896
	b	2-66	185.6	1.38	134.6	1.188	0.404	808
Highlandcroft Plutonic Series								
HHF-1	w_I	5-65	45.60	47.79	0.95	0.710	0.002	357
FSM-1	w_I	5-65	51.05	14.76	3.46	0.728	0.027	1296
FSM-4	w_I	6-65	48.87	12.26	3.99	0.732	0.033	1380
Ammonoosuc Volcanics								
AWR-1	w_I	7-65	3.52	22.10	0.16	0.709	0.001	1419
AWR-2	w_I	6-65	6.89	9.03	0.76	0.711	0.005	979
AWR-6	w_I	6-65	8.01	6.39	1.25	0.709	0.001	199
ASH-1	w_I	7-66	20.34	5.99	3.40	0.726	0.025	1187
ASH-2	w_I	1-66	22.60	5.33	4.24	0.733	0.035	1356
ASH-3	w_{III}	1-66	17.59	6.22	2.83	0.725	0.023	1360

EXPLANATION

(1) $$T = \frac{1}{\lambda} \ln\left(\frac{\left(\frac{Sr^{87}}{Sr^{86}}\right) - 0.708}{\left(\frac{Rb^{87}}{Sr^{86}}\right)} + 1 \right)$$

1 year = 3.1557×10^{7} sec.
$\lambda Rb^{87} = 1.39 \times 10^{-11}$ year^{-1}

(2) determined by Xray-fluorescence spectrometry

(3) calculated from spiked sample except where noted
normalized to $Sr^{86}/Sr^{88} = 0.1194$

(4) measured on un-spiked aliquot
normalized to $Sr^{86}/Sr^{88} = 0.1194$

(5) subscripts denote replicate analyses of splits of powdered sample

PHASES

- w - whole-rock
- ap - apatite
- p - plagioclase
- k - microcline
- m - muscovite
- b - biotite

Table 2, concluded

possible variation of $\pm$ 2 percent in the Rb/Sr ratios (indicated by dimensions of the crosses in the figures) and uncertainty of $\pm$ 0.002 in the Sr^{87}/Sr^{86} ratio of the initial Sr. In most cases the uncertainty in the composition of common Sr produces most of the error.

Each of the figures shows one or more points which are displaced from the isochrons beyond the estimated limits of analytical uncertainty. The analyses of some of these points were repeated (dissolution of replicate splits of the powdered sample), but in no case did such replication result in significantly better fit. The discrepancies are discussed individually, but in most cases it has not been possible to isolate the causes unequivocally, so the discrepancies add to the overall uncertainty of the age determinations. Most, but not all, of these effects trended in the direction of low apparent ages.

Granitic Core-Rocks of the Lebanon Dome

The conditions for Rb-Sr whole-rock dating are best satisfied by rocks from the core of the Lebanon Dome, data for which are plotted in Figure 5. Sample LCW-1, fine-grained granite, appears to be younger than the other rocks, and no line passing through the points for this sample provides a satisfactory fit to the other data. If the other samples are assumed to be cogenetic, their data are fitted reasonably by the solid line which has a slope corresponding to an age of 440 $\pm$ 20 m.y. ($\lambda\ Rb^{87} = 1.39 \times 10^{-11} year^{-1}$) and an initial ratio, $(Sr^{87}/Sr^{86})_0 = 0.706$. The slope of this line is strongly controlled

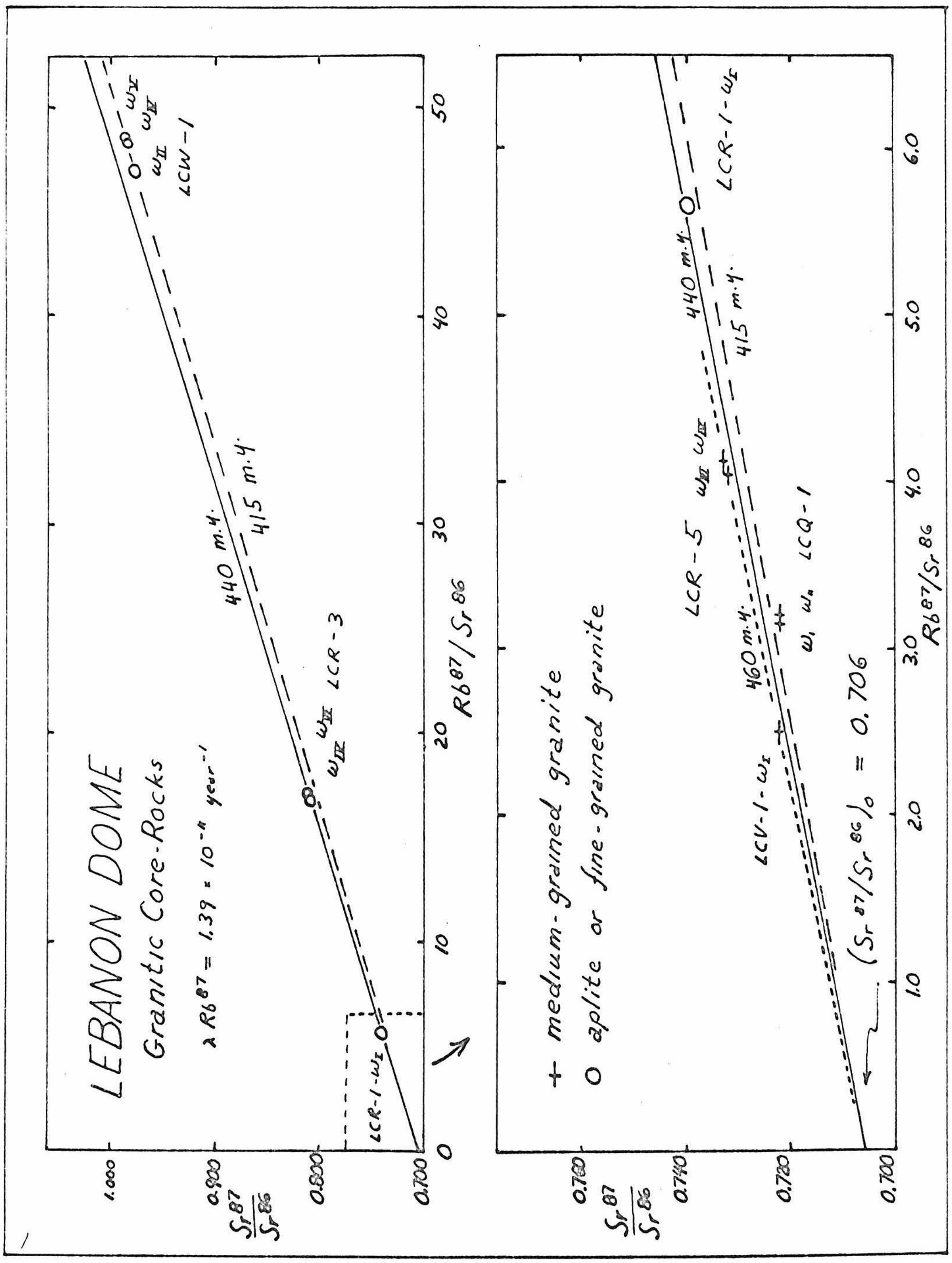


Figure 5. Sr-evolution diagram for whole-rock samples from the Lebanon Dome, New Hampshire.

by the aplite samples LCR-1 and LCR-3 , and a slightly different interpretation is possible if the aplites are considered separately from the medium-grained granites LCR-5, LCV-1, and LCQ-1 which are the most common constituents of the Lebanon Dome. A line connecting the data for samples LCR-5 and LCV-1 has a slope of 460 $\pm$ 30 m.y. and an initial ratio, $(Sr^{87}/Sr^{86})_0$ = 0.706. Any line which assigns equal weight to the data for sample LCQ-1 yields an initial ratio, $(Sr^{87}/Sr^{86})_0$, less than 0.703 which is probably an unreasonably low value for granitic rocks of this age (Hedge and Walthall, 1963, p. 1216). Sample LCQ-1 is in all respects typical of the granitic rocks of the core of the Lebanon Dome and there is no reason to believe it is younger than the other granitic rocks. Its apparent age may have been lowered by some form of disturbance, probably during one of the periods of subsequent metamorphism.

The dashed line connecting $(Sr^{87}/Sr^{86})_0$ = 0.706 and the points for sample LCW-1 (fine-grained granite) corresponds to an age of 413 $\pm$ 15 m.y. The line is not a good fit to the other data and it appears that sample LCW-1 is either younger than the other samples or has had its apparent age reduced by disturbance during metamorphism. For the sample to have an undisturbed age of 440 m.y. its initial Sr^{87}/Sr^{86} composition would have to be 0.680. This is lower than the value accepted for primordial Sr and is unreasonable. The sample is sufficiently radiogenic that an uncertainty of 0.003 in the value of $(Sr^{87}/Sr^{86})_0$ affects the apparent age by only 5 m.y.

To summarize, the fine-grained granite has an apparent age of 413 ± 15 m.y., but the apparent age may be low due to disturbance during metamorphism. The aplite samples, LCR-1 and LCR-3, have apparent ages of 440 ± 20 m.y., whereas the medium-grained granite samples, LCR-5 and LCV-1, have apparent ages of 460 ± 30 m.y. The difference in the ages of the aplite and medium-grained samples may not be significant.

Granitic Core-Rocks of the Mascoma Dome

The geochronologic results for the granitic core-rocks of the Mascoma Dome are similar to those discussed for the Lebanon Dome, although the Mascoma samples are less radiogenic and the errors are correspondingly greater. Rb-Sr whole-rock samples MJH-1, MMQ-3, and MMQ-11 lie on an isochron with an initial ratio, $(Sr87/Sr86)_0 = 0.705$, and a slope corresponding to an age of 440 ± 45 m.y. (Figure 6). The error is strongly affected by uncertainty in the composition of initial Sr. The data for sample MMS-2 lie significantly above this line. Samples MMS-2 and MMQ-11 are lithologically similar quartz monzonite samples, both collected in what appears to be a single pluton. They were collected about 1½ miles apart. If both samples are cogenetic closed systems, sample MMS-2 would appear to have incorporated initial Sr 0.4 percent more radiogenic than that incorporated by sample MMQ-11. The Rb-Sr data suggest that aplite (MMQ-3) and medium-grained quartz monzonite (MMQ-11) from a single locality are cogenetic and have common Sr of the same isotopic composition. If the spread in the data is due to variations in the composition of initial Sr

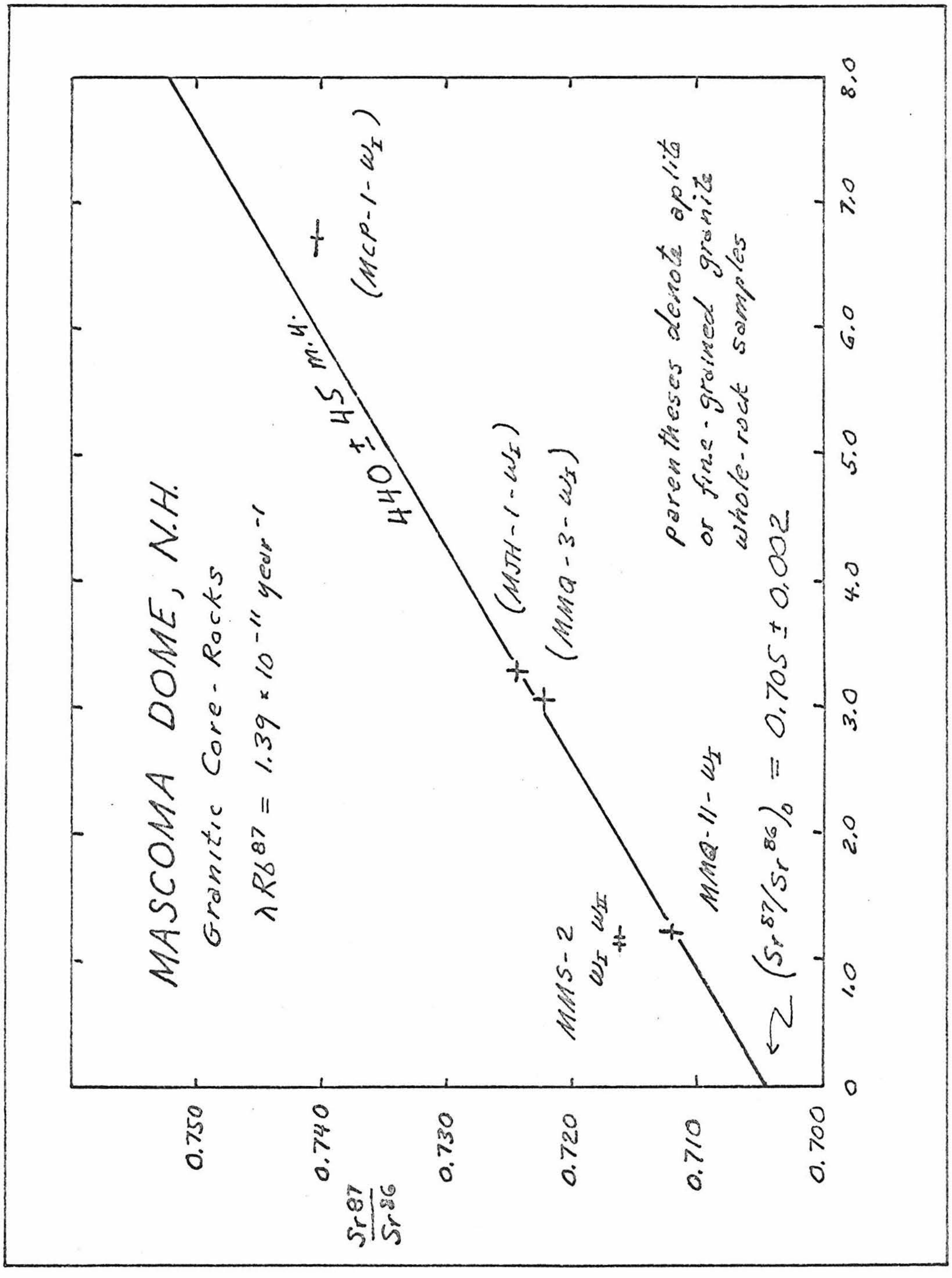


Figure 6. Sr-evolution diagram for whole-rock samples from the Mascoma Dome, New Hampshire.

within the Mascoma Dome it is likely that aplite collected near sample MMS-2 would show initial Sr of the composition indicated for the medium-grained quartz monzonite (MMS-2) at the same locality. The author plans to search for aplite at this locality to test this hypothesis.

Whole-rock sample MCP-1, from a body of fine-grained granite in the southern part of the Mascoma Dome has an apparent age of 370 $\pm$ 15 m.y. (assuming initial Sr87/Sr86 = 0.705). The locality lies near the sheared zone of a Triassic (?) fault and the low apparent age may be due to disturbance. Alternately the granite may be a small body younger than the main mass.

Highlandcroft Series

The author attempted to date several samples of rocks from the Highlandcroft Series for comparison with the Oliverian Dome results, but was unable to find samples with a satisfactory spread in Rb/Sr ratios from any single pluton. Two samples, FSM-4 and FSM-1, from the Fairlee Pluton about ten miles north of Hanover, New Hampshire (see Figure 1), are sufficiently radiogenic for dating but do not provide sufficient basis for estimating the composition of initial Sr. If $(Sr87/Sr86)_0$ is assumed to be 0.706 $\pm$ 0.002 the samples have an apparent age of 450 $\pm$ 40 m.y. (Figure 7). A line passed through the data for these samples and sample HHF-1 (from the Highlandcroft Pluton west of Littleton, New Hampshire) has a slope of 540 m.y. and an initial intercept at 0.702. This initial ratio is low for granitic rocks of this age (Hedge and Walthall, 1963, p. 1216), and the

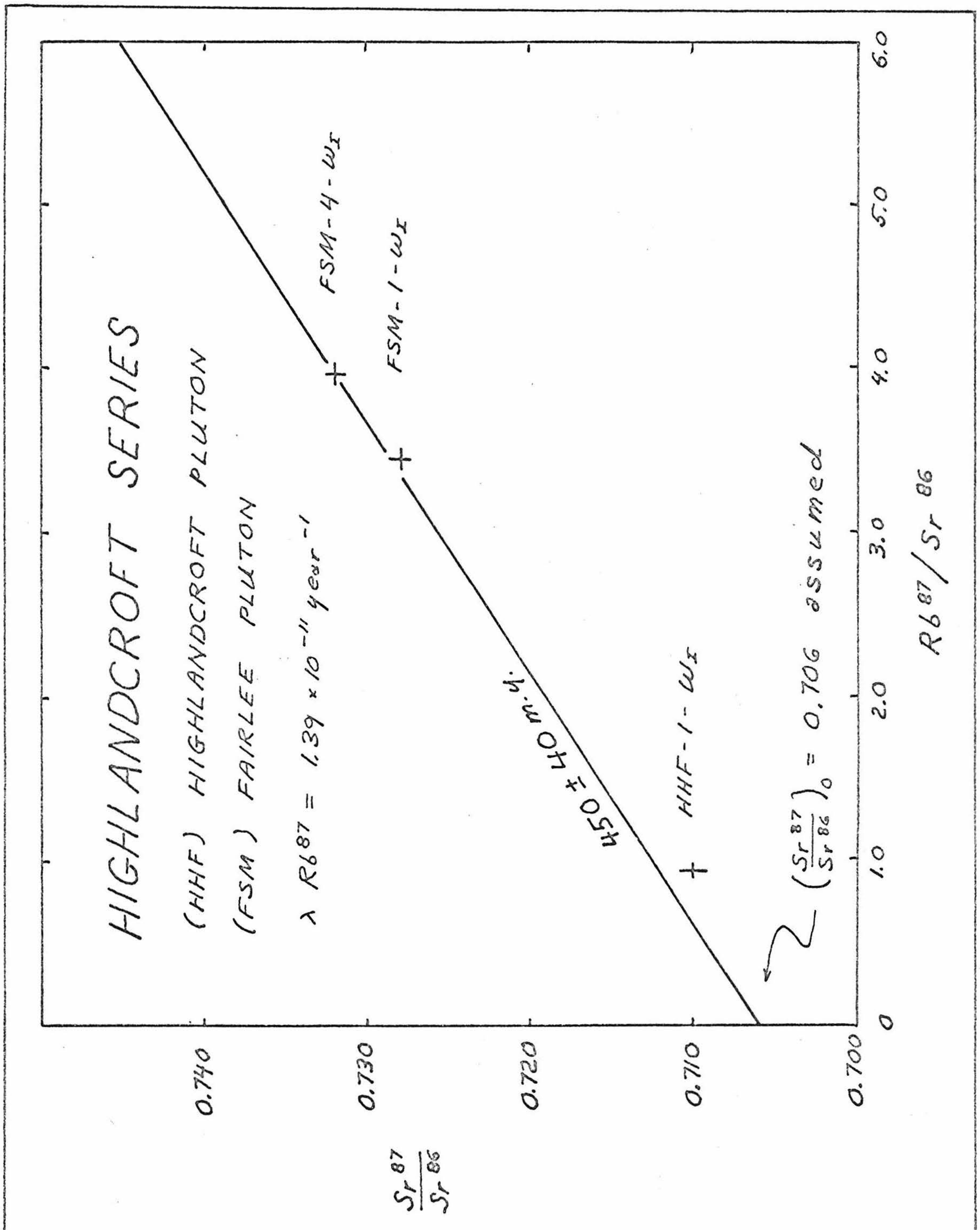


Figure 7. Sr-evolution diagram for whole-rock samples from the Highlandcroft Series.

apparent age is probably anomalous. Much more data are needed to estimate possible variations in age, composition of initial Sr, and response to disturbance among plutons of the Highlandcroft group.

Ammonoosuc Volcanics

The Ammonoosuc Volcanics flanking the Mascoma Dome yielded no samples sufficiently Rb-rich to be satisfactory for Rb-Sr whole-rock dating. Marginally satisfactory samples were obtained from exposures of the Ammonoosuc Volcanics west and south of Littleton, New Hampshire. These samples come from discontinuous belts of exposures which cannot be traced directly to the continuous belt of Ammonoosuc Volcanics which mantles the Oliverian Domes. Correlation of these belts has been discussed by Billings (1937, p. 495-499). The data are plotted in Figure 8. A line with a slope corresponding to an age of 440 $\pm$ 30 m.y. and an intercept of $(Sr87/Sr86)_0 = 0.707 \pm 0.002$ is the best fit to five of the data-points, but a sixth sample, AWR-6, lies far below the line. Sample AWR-6 contains a small amount of carbonate which is soluble in dilute HCl and the sample may not be a closed system with respect to Sr. The circles with error-brackets plotted on Figure 8 are the data from which Brookins and Hurley (1965, p. 11) deduced an age of 418 $\pm$ m.y. for the Ammonoosuc Volcanics. Sample R5069 is from the same outcrop as all the samples labelled ASH, and the other Brookins and Hurley samples come from nearby. In light of the scatter in the data the present author suggests that the $\pm$ 30 m.y. estimated uncertainty is more appropriate than the $\pm$ 15 m.y. uncertainty quoted by Brookins and Hurley. The 440 m.y. line appears to be a better fit to the data.

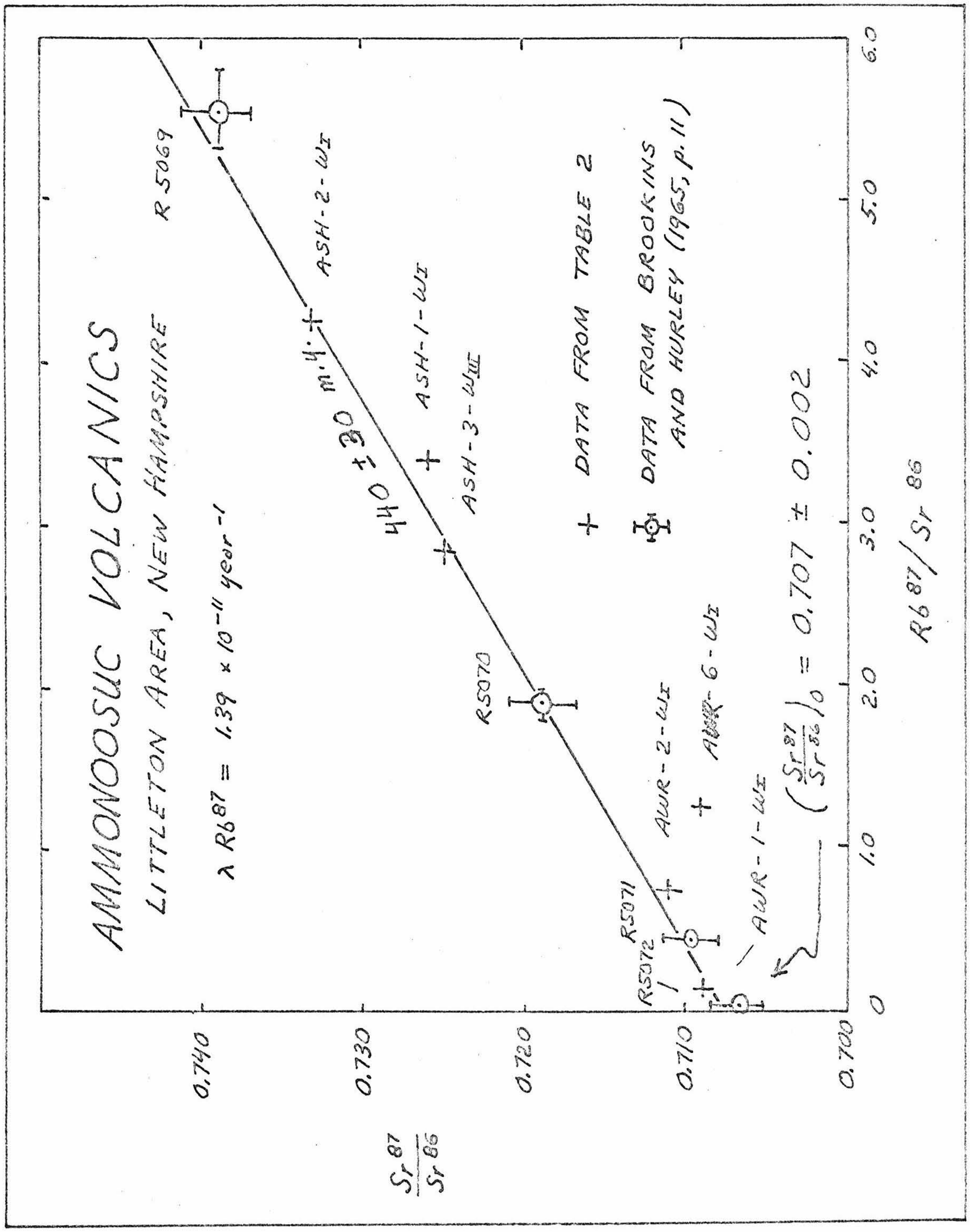


Figure 8. Sr-evolution diagram for whole-rock samples of the Ammonoosuc Volcanics, Littleton area, New Hampshire.

Determinations on Mineral Separates

Rb-Sr Data

Rb-Sr data on minerals separated from the granitic core-rocks of the Lebanon and Mascoma Dome are tabulated in Table 2 and plotted in Figures 9 and 10. The treatment of the Sr data is the same as for the whole-rock samples previously discussed. When rocks are disturbed by metamorphism it is possible for isotopes of Rb and Sr to be exchanged over distances which are large compared to the dimensions of the mineral grains. If this exchange tends towards isotopic equilibrium the composition of Sr in each grain approaches the value characteristic of the rock. If the minerals are closed to migration of Rb and Sr subsequent to the disturbance their data will plot as a linear array on the strontium evolution diagram. The slope corresponds to the time elapsed since the disturbance and the intercept indicates the Sr^{87}/Sr^{86} ratio in the rock at the time of the disturbance. If the scale of migration of Rb and Sr isotopes is small compared to the scale over which the rock samples are homogeneous the whole-rock isochrons will be unaffected by the disturbance. If there are multiple disturbances, each of which drives the minerals to isotopic equilibrium, the isochron technique detects only the last such event, and evidence for the early events is lost from the mineral isochrons. The technique of interpreting mineral isochrons is further discussed by Lanphere, and others (1964).

Minerals were separated from sample MJH-1, fine-grained granite from the Mascoma Dome (Figure 9), and from sample LCR-5, medium-grained

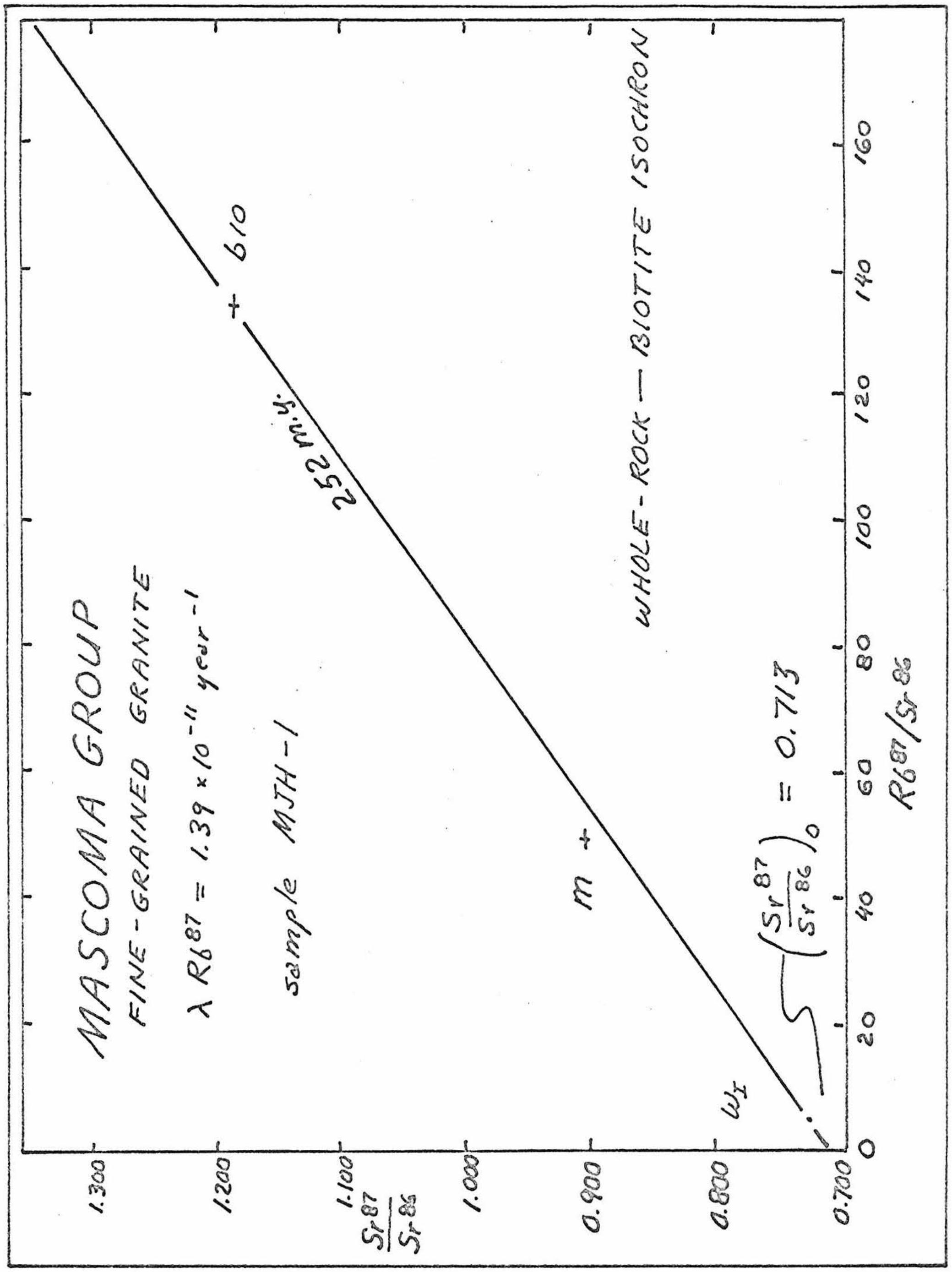


Figure 9. Sr-evolution diagram for sample MJH-1, Mascoma Dome, New Hampshire.

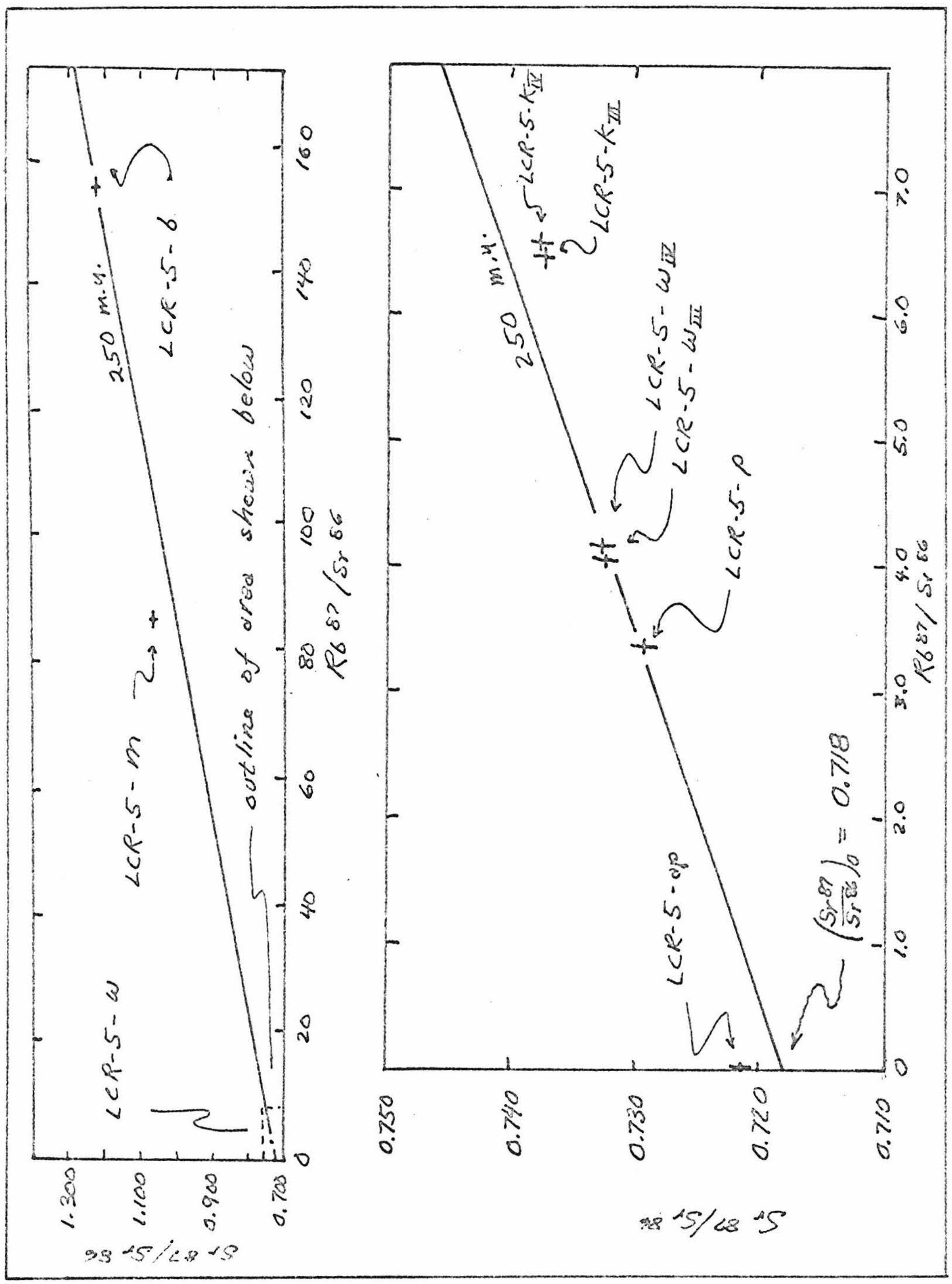


Figure 10. Sr-evolution diagram for sample LCR-5, Lebanon Dome, New Hampshire.

granite from the Lebanon Dome. The whole-rock-biotite isochrons for both rocks indicate ages of 250 $\pm$ 5 m.y. In sample LCR-5 other minerals lie close to this isochron suggesting approach towards isotopic equilibration during a disturbance at about 250 m.y. The presence of excess radiogenic Sr in apatite and plagioclase is strong evidence for migration of Sr isotopes, but the role of Rb migration is more difficult to ascertain. A tendency for muscovite to plot above the whole-rock biotite isochron is noted in both samples and has been observed in other samples from New England.

U-Pb Determinations on Zircon Separates

Isotopic studies on zircons have provided useful information on the geochronology of several samples from the Mascoma Dome. The analytical data are tabulated in Table 3 and are plotted on a Pb^{206}/U^{238} - Pb^{207}/U^{235} diagram in Figure 11. The analytical techniques are discussed in Appendix 1 and the samples are described in Appendix 2. The interpretations which follow are based chiefly on the Pb^{207}/Pb^{206} ratios of the zircons analyzed. A precision error of $\pm$ 0.5 percent in this ratio is indicated by the size of the circles in Figure 11. At the time of this writing the author's experiments to determine the estimated precision error in the ratios Pb^{206}/U^{238} and Pb^{207}/U^{235} are incomplete, and for this reason interpretations of the type which depend critically on the precision of these ratios are not treated in the discussion which follows. Zircons have been analyzed from samples of Holts Ledge Gneiss (MHL - 1), and both fine-grained granite (MJH - 1) and medium-grained quartz monzonite (MMQ - 11) of the Mascoma Group. Zircons from sample MMQ - 11 were fractionated to yield sub-populations which differ in size and radioactivity. These fractions are believed to be cogenetic, but show differential response to disturbance. These fractions are discussed first.

The zircons separated from sample MMQ - 11 show a wide range of chemical and physical properties. There is no practical method for extracting all of the zircon from a sample of rock, and there is no reason to expect that the zircons lost have the same properties as the

Sample	concentration U^{233} (10^{-3} moles/g)	concentration Pb^{206} (10^{-3} moles/g)	observed relative isotopic abundances (1) Pb^{204}	Pb^{206}	Pb^{207}	Pb^{208}	corrected ratios (2) $\frac{Pb^{206*}}{U^{233}}$	$\frac{Pb^{207*}}{U^{235}}$	$\frac{Pb^{207*}}{Pb^{206*}}$	calculated ages $\frac{Pb^{206*}}{U^{238}}$	$\frac{Pb^{207*}}{U^{235}}$	$\frac{Pb^{207*}}{Pb^{206*}}$
HOLTS LEDGE GNEISS												
MHL-1 Zr-9	363	23.72	0.0306	100.00	5.992	17.73	0.05636	0.4988	0.05542 ±0.0003	412	416	440 ± 25
MASCOMA GROUP, FINE-GRAINED GRANITE												
MJH-1 Zr-9	515	30.81	0.0468	100.00	6.192	11.37	0.05982	0.4533	0.05504 ±0.0003	378	384	425 ± 25
MASCOMA GROUP, MEDIUM GRAINED QUARTZ MONZONITE												
MMQ-11 Zr-9	601	37.53	0.0533	100.00	6.321	18.43	0.06245	0.4763	0.05538 ±0.0003	394	401	439 ± 25
Zr-mm	404	24.30	0.09161 (3)	100.00	7.047 (3)	19.25 (3)	0.0602	0.4630	0.05584 ±0.0003	380	392	457 ± 35
Zr-mj	1018	52.98	0.0755	100.00	6.671	16.78	0.05204	0.3987	0.05564 ±0.0003	330	345	449 ± 25

Notes
(1) corrected for scale factor and square root of mass ratio
(2) * denotes radiogenic; corrected for common Pb 204 : 206 : 207 = 1.0 : 18.3 : 15.7, mid-Paleozoic common Pb (Doe and others, 1965, p. 1954); U^{233}/U^{235} = 137.7
(3) calculated from spiked aliquot; Pb^{208} determined by extrapolation from co-fractions

Table 3. U-Pb analytical data for zircon samples from the Mascoma Dome, New Hampshire.

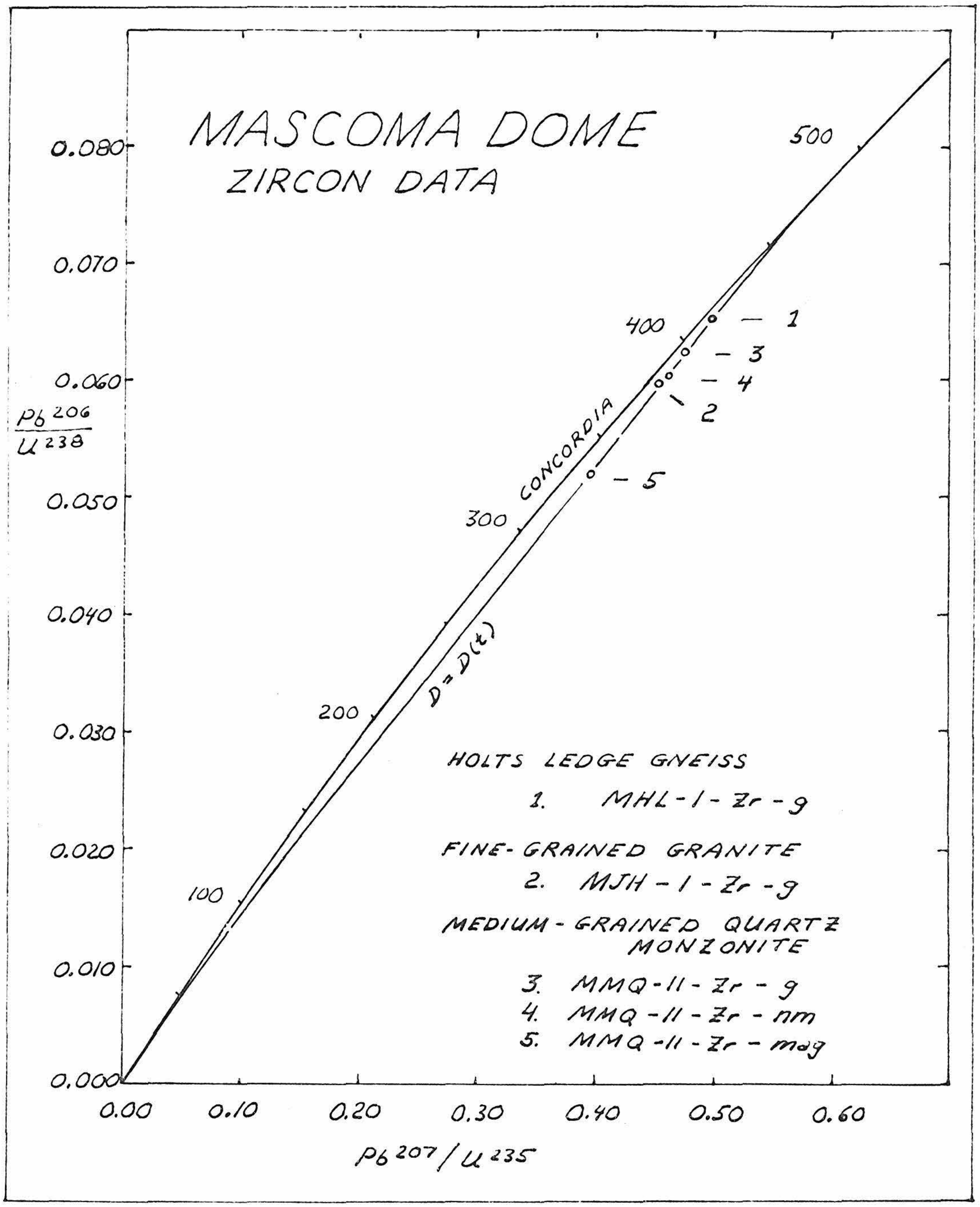


Figure 11. Pb^{206}/U^{238} -- Pb^{207}/U^{235} diagram for zircon samples from the Mascoma Dome, New Hampshire.

zircons recovered. All of the zircon samples are called "fractions" to emphasize the presence of this inevitable fractionation. About half of the heavy-mineral concentrate was purified to yield a population of zircons referred to as the gross fraction (MMQ-11-Zr-g). The other half of the concentrate was purified at a later date and was fractionated on the Franz separator to yield the following fractions: most-magnetic (MMQ-11-Zr-mag), intermediate (MMQ-11-Zr-i), and least-magnetic (MMQ-11-Zr-nm). Because the two halves of the heavy-mineral concentrate were treated separately it is not possible to say to what degree the gross fraction (g) is representative of the population from which the other fractions were separated. The gross fraction (g) is a heterogeneous population of irregular, stubby zircons with mostly very poor development of the crystal faces. Many of the zircons are cloudy and dark brown, particularly in their centers. Dark inclusions are common, but distinct zones, cores, and overgrowths are rare. The more-magnetic fraction (mag) is strongly enriched in small, stubby, very poorly-formed, cloudy, dark-brown zircons. The less-magnetic fraction (nm) is enriched in small, clear, pale-straw-colored zircons which are nearly free of inclusions and which show some tendency towards better form. The intermediate fraction (i) contains zircons which are mostly larger than those in the two extreme fractions. Table 3 indicates that the magnetic separation was successful in concentrating fractions which differ significantly in radioactivity.

Numerous models have been proposed to account for the systematics of discordant zircon data (Wetherill, 1956, and 1963; Tilton, 1960;

Silver and Deutsch, 1963; Wasserburg, 1963; Steiger and Wasserburg, 1966; and Allegre, 1967). Most of these papers discuss the interpretation of discordant zircon data in terms of coupled parent-daughter diagrams of which the Pb206/U238 - Pb207/U235 diagram in Figure 11 is an example. In some cases it has been possible to demonstrate that one particular effect was predominant in producing the discordance (eg. Silver and Deutsch (1963) showed that discordance of zircons from the Johnny Lyon granodiorite is best explained by episodic loss of Pb), but there are other cases where the mechanism is not clear. In rocks with complex thermal histories it is likely that several discordance-producing-mechanisms may act in concert. Understanding of the mechanism is critical if conclusions regarding the nature and timing of disturbing events are to be drawn from the data, but such understanding is less critical in interpreting the primary ages of samples. Wetherill (1963) and Wasserburg (1963) observed that a wide class of such mechanisms result in a linear array of data in the neighborhood of the primary intercept with the concordia curve (the locus of data for concordant samples), and Wasserburg (1963, p. 4830) demonstrated that for data in this region of the diagram, determination of the primary intercept on concordia is only slightly affected by uncertainty in the mechanism which produced the discordance.

Three zircon fractions (g, mag, and nm) from sample MMQ-11 are plotted on a Pb206/U238 - Pb207/U235 diagram (Figure 11). Within experimental error the three samples have the same Pb207/Pb206 ratio, 0.05565 ± 0.00030. This corresponds to a primary age of 450 ± 25 m.y. "Radiation-damage diffusion curves" with primary intercepts between

450 and 460 m.y. (Wasserburg, 1963, p. 4830) fit the data within the experimental error, although there is not sufficient data to establish that such a mechanism is truly responsible for the discordance. As discussed in the preceding paragraph the interpretation of the primary intercept would not be significantly affected by choice of any of the other discordance models. The sample was probably disturbed at about 360 m.y. (the estimated time of middle-grade metamorphism in the Mascoma area), about 250 m.y. (the disturbance recorded by the Rb-Sr mineral isochrons), and possibly again about 180 m.y. (the time of emplacement of the White Mountain Magma Series nearby). The nature and duration of these disturbances are not well understood, and it is not known whether they are discrete episodes. It is interesting to note that none of these disturbances is recorded in any simple way by the zircon data plotted in Figure 11. The author hopes that refinement of the concentration data will provide more information on the nature of the disturbances, but does not expect that the primary intercept will be altered significantly.

It is necessary to discuss the probable origin of the zircons obtained from sample MMQ-11. The heterogeneity of the population raises the question as to whether the sample consists of a mixture of zircons with significantly older than 450 m.y. If older zircons are present it would be highly fortuitous for them to be allocated among the various fractions in precisely such a way as to produce the observed relationships. If the fractions are assumed to be mixtures containing old zircons of a given age, the age of the younger zircons must be different in each of the fractions, and again the linear arrangement would be fortuitous. The differences exhibited probably result from differential response to disturbance of a heterogeneous but essentially cogenetic suite of zircons.

If the quartz monzonite formed from rock containing zircons significantly older than 450 m.y. all traces of the older age were destroyed during formation. As in the case of mixing, even small quantities of older zircon would upset the linear relationship of the data, barring a fortuitous partition of the zircons among the fractions. The arguments for both mixing and discordance become more flexible as the age of the older component of zircons approaches 450 m.y. Due to the shallow curvature of concordia in the region of interest the possibility that the samples contain some zircons as old as 500 m.y. cannot be eliminated with the present data.

Heavy mineral concentrates were purified to yield gross fractions of zircons from samples MHL-1 (Holts Ledge Gneiss) and MJH-1 (fine grained granite of the Mascoma Group). Within experimental error these two samples lie on the same diffusion curve as the zircons from sample MMQ-11. Within experimental error fraction (g) from sample MHL-1 has the same primary age as the fractions from sample MMQ-11. The Pb207/Pb206 ratio of the zircons from sample MJH-1 is lower than for the other samples, and corresponds to a primary age of 425 $\pm$ 25 m.y. This difference may not be analytically significant, but its trend is consistent with the observation that some of the fine-grained granite bodies yield apparent Rb-Sr whole-rock ages which are lower than the apparent ages of the coarser granitic rocks.

DISCUSSION

The cores of most of the Oliverian Domes contain rocks which exhibit a wide range of geological properties and relationships. While recognizing this variety, most previous geologists have not tried to subdivide the core-rocks when discussing their origins. Attempts to find a single mode of origin which "explains" the wide range of features observed have not been successful. Many theories for the origin of the domes--some quite complex and sophisticated--have been advanced, but none has proved wholly satisfactory. The author's approach has been to recognize that the diverse properties of the core-rocks may reflect a diversity of origin. In the discussion which follows the author will rely on a plausible subdivision of the core rocks in proposing an origin for the Mascoma Dome which appears consistent with the geologic and isotopic data presented earlier in this paper. The other Oliverian Domes, particularly those in New Hampshire, are similar in most respects to the Mascoma Dome and probably originated in much the same manner.

Age and Origin of the Holts Ledge Gneiss

The Holts Ledge Gneiss is the most abundant but least understood rock unit exposed in the Mascoma Area. Most previous geologists (eg. Chapman, 1939; Hadley, 1942) have concluded that the Holts Ledge Gneiss is Devonian intrusive rock, but this conclusion has not been universally accepted. Eskola (1949, p. 470-471) suggested that the core-rocks of the Oliverian Domes (which include rocks assigned here to the Holts Ledge Gneiss) are older than the (Ordovician?) Ammonoosuc Volcanics, and several geologists (eg. Robinson, 1963) have suggested these rocks might locally be as old as Precambrian. Other geologists (eg. J. B. Thompson, Jr., oral communication, 1964) have inferred these rocks are in part "metamorphosed Ammonoosuc Volcanics" but this inference has never been supported in print. The author will attempt to show that: (1) the Holts Ledge Gneiss is a distinct unit separate from the Ammonoosuc Volcanics and the Mascoma Group; (2) the stratification of the gneiss is not the result of deformation at the time the dome was formed; (3) the Holts Ledge Gneiss is not intrusive into the Ammonoosuc Volcanics or younger units, but probably lies stratigraphically beneath them; (4) the gneiss is not a strongly metamorphosed or remobilized Precambrian formation; and (5) an Ordovician (?) volcanic origin for the protolith of the gneiss is plausible but not conclusively demonstrated. Rocks similar to the Holts Ledge Gneiss are common in most of the Oliverian Domes and probably originated in much the same manner.

Figure 12 is a plot of the relative normative abundance of quartz, plagioclase, potassium feldspar, and corundum. A normative presentation was chosen as a means of including the variable amounts of Si, Al, Ca, and K in biotite and epidote which are minor but variable constituents of most of the rocks. Most of the points (open symbols) are molecular norms calculated from thin-section modes and the estimated composition of the minerals. A few points (solid symbols) are molecular norms calculated from published chemical analyses. The abundance of corundum is indicated by the length of the slashes. The procedure is fairly reliable for the elements plotted on the

Figure 12 Explanation. The triangular diagrams plot the relative abundance of normative quartz (Qz), plagioclase (Ab + An), and potassium feldspar (Or). Normative corundum in excess of 2 percent is indicated by the length of the vertical slashes. Open symbols represent norms calculated from thin-section modes and estimated compositions of the minerals; solid symbols represent norms calculated from published chemical analyses. The data are from the following sources: circles, Chapman and Schweitzer (1947); upright triangles, Hadley (1942); inverted triangles, Chapman (1939); diamonds, Billings (1937); and squares, thin-sections examined by the present author. Diagrams d and e are replotted from Hopson (1964, p. 34). The solid circles are samples of altered marine volcanic siltstone and sandstone from the Eocene Ohanapecosh Formation, Washington. Diagram e plots the average composition of the following rock-types: shale (SH), arkose (AK), sub-graywacke (SG), graywacke (GW), rhyolite (R), dellenite (DL), rhyodacite (RD), dacite (D), andesite (A), and tholeiitic basalt (B).

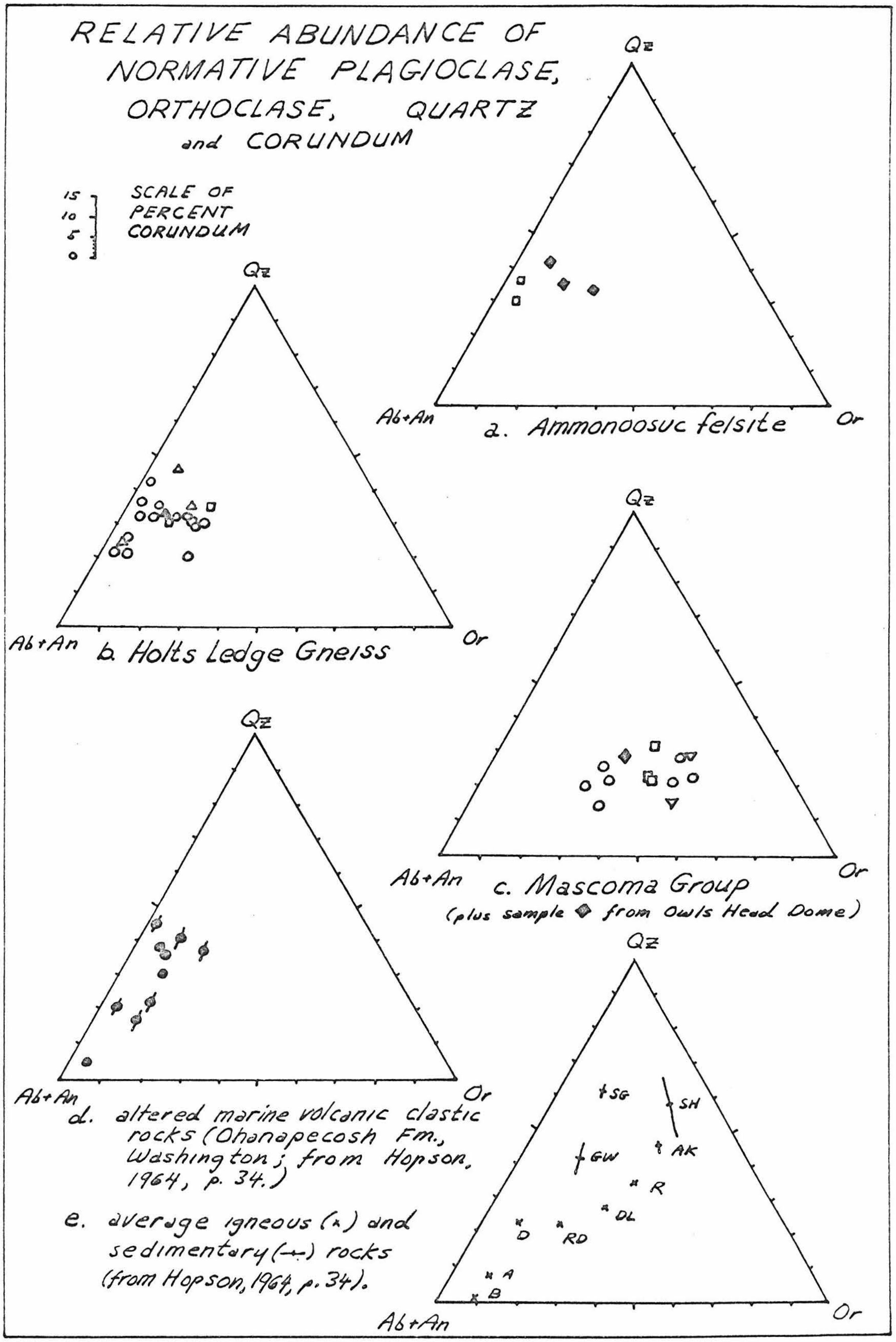

RELATIVE ABUNDANCE OF NORMATIVE PLAGIOCLASE, ORTHOCLASE, QUARTZ and CORUNDUM
15
10
5
0
SCALE OF PERCENT CORUNDUM
Qz
Ab+An
Or
a. Ammonoosuc felsite
b. Holts Ledge Gneiss
c. Mascoma Group
(plus sample ◆ from Owls Head Dome)
d. altered marine volcanic clastic rocks (Ohanapecosh Fm., Washington; from Hopson, 1964, p. 34.)
e. average igneous (x) and sedimentary (+) rocks (from Hopson, 1964, p. 34).
SG
SH
GW
AK
R
DL
D
RD
A
B

diagram and serves to point out the gross differences and similarities among the rock units. Because Ca and Na are summed on the diagram, only second-order errors are introduced by uncertainty in the composition of plagioclase. Since biotite and epidote are present in minor amounts, uncertainty in their composition also introduces only second-order errors in the elements plotted.

Figure 12 shows that some of the felsite layers in the Ammonoosuc Volcanics are similar in composition to some of the light-colored gneiss layers in the Holts Ledge Gneiss. This similarity was noted by Page (1940) who, as discussed previously, included about a thousand feet of rock similar to the Holts Ledge Gneiss in the Ammonoosuc Volcanics south of the Owls Head Dome (see Figure 1). The two units can be readily distinguished, however, by the scarcity of amphibolite and massive character of the Holts Ledge Gneiss as compared with the Ammonoosuc Volcanics. Figure 12 shows that the average composition of the Holts Ledge Gneiss is quite different from that of the Mascoma Group, the only exception being rare layers of rhyolitic felsite (not plotted on Figure 12) occurring near the top of the Holts Ledge Gneiss. As discussed previously, where exposures are adequate the contact between the Holts Ledge Gneiss and the Mascoma Group is sharp and can be readily mapped.

Severe deformation, probably related to the formation of the dome, is evident near the boundary between the core and the mantle. Regarding the Clough Formation, Hadley (1942, p. 161-162) wrote:

> On the northwest flank of the Mascoma Dome, where the deformation seems to have been greatest, ellipsoidal pebbles have axial ratios between 1:2:4 and 1:3:8, the shortest axis always normal to the bedding and schistosity. Assuming them to have had originally the high degree of sphericity characteristic of well-rounded quartz pebbles . . . they have been flattened normal to the bedding by at least 35 percent, and elongated nearly 50 percent. Thousands of pebbles, flattened and elongated all in the same direction, attest a severe, plastic thinning and stretching of the walls of the domes.

The present author does not know whether this deformation has tended to enhance or efface the stratified appearance of the Holts Ledge Gneiss, but several lines of evidence suggest the stratification was not created by this deformation. Most importantly, the layering in the gneiss does not extend into the granitic rocks, even where the granite lies in the zone of probable maximum deformation near the core-mantle boundary. This suggests that the contrast between the massive, homogeneous granitic rocks and the more heterogeneous, weakly statified Holts Ledge Gneiss was not established by the deformation which formed the dome and stretched cobbles in the Clough Formation. In the Chester Dome (not an Oliverian Dome; see Appendix 4) compositional banding appears to have formed by a "streaking out" of older dark-colored layers in planes of shear parallel to the dome margins. Layers in the Holts Ledge Gneiss are thicker and have greater lateral persistence than the secondary banding in the Chester Dome. The particular lithological contrasts and discrete planes of separation which mark layers in the Holts Ledge Gneiss (see previous discussion of geology) resemble bedding and were probably not formed by deformation. If the layering were secondary banding produced by the dome-forming deformation some traces of the original layering might be preserved. Such traces are preserved in the Chester Dome (see Appendix 4) and their absence in the less-deformed Mascoma Dome is a further argument against a secondary origin for the layering.

The constant stratigraphic position and concordance of the Holts Ledge Gneiss suggest more that the gneiss lies stratigraphically beneath the Ammonoosuc Volcanics than that it is intrusive into them. Moreover there is no positive evidence which requires that the gneiss be considered intrusive. Previous geologists (eg. Chapman, 1939; and Hadley, 1942) concluded that the rocks assigned to the Holts Ledge Gneiss are intrusive chiefly on the basis of their igneous composition and an inferred relationship to the granitic rocks which locally appear intrusive. Most of the structural arguments formerly advanced in support of this conclusion have subsequently been given alternate explanations. The igneous composition of the gneiss can be explained by a volcanic as well as an intrusive origin. The differences shown in Figure 12 indicate the unlikelihood that the Holts Ledge Gneiss and the granitic rocks of the Mascoma Group are unaltered members of a single differentiated magma series. The trend of decreasing normative quartz with increasing normative potassium feldspar is not a common differentiation trend. A small group of rocks assigned to the Mascoma Group have the composition of quartz diorite, and are lithologically similar to parts of the Holts Ledge Gneiss, but as discussed previously they crosscut the upper part of the Holts Ledge Gneiss (see Figure 2). As discussed previously the granitic rocks of the Mascoma Group also crosscut the Holts Ledge Gneiss. Because the rocks of the Mascoma

Group are younger than the Holts Ledge Gneiss and not closely related to it chemically, the crosscutting relationships of the Mascoma Group do not constitute an argument that the Holts Ledge Gneiss is intrusive. Nowhere has the author seen rocks assigned to the Holts Ledge Gneiss crosscut the Ammonoosuc Volcanics or younger formations. As shown in Figures 2, 3, and 4 the upper contact of the Holts Ledge Gneiss is parallel with the base of the Ammonoosuc Volcanics. Nowhere along the six miles of contact mapped does the gneiss cut across the thin felsite zone near the base of the Ammonoosuc Volcanics. Along most of the 250-mile length of the Oliverian Dome belt, rocks similar to the Holts Ledge Gneiss are exposed at essentially the same stratigraphic position beneath the Ammonoosuc Volcanics. In the absence of positive contrary evidence, these relationships strongly imply that the Holts Ledge Gneiss lies stratigraphically beneath the Ammonoosuc Volcanics.

The Holts Ledge Gneiss retains no presently-detectable traces of a geologic history prior to 450 to 500 m.y. Zircons were analyzed from a sample (MHL-1) of typical granodioritic gneiss from near the base of the formation in the type area. The population of zircons analyzed appears to contain no significant quantity of zircons with Pb207/Pb206 ages greater than 500 m.y. Such zircons would have to be "counter-balanced" with other zircons having Pb207/Pb206 ages less than 450 m.y. and it would be fortuitous for the total fraction analyzed to lie as it does on the same 450 $\pm$ 25 m.y. chord as the zircon samples from the Mascoma Group.

Experience with other gneiss domes indicates that older rocks retain traces of evidence indicative of their ancestry through very high degrees of metamorphism and deformation. In the Baltimore (Maryland) Gneiss Domes Precambrian core-rocks have been metamorphosed and deformed at kyanite-staurolite grade (Hopson, 1964, p.78), yet their zircons retain clearcut evidence for the Precambrian age of the rocks (Tilton, Wetherill, Davis, and Hopson, 1958). Thompson (1950) has shown that probable Precambrian core-rocks in the Chester Dome of southeastern Vermont were affected by kyanite-staurolite grade Paleozoic metamorphism. Thompson (1950, p. 68) states that the mantle-rocks flanking the Chester Dome have been reduced in thickness by a factor of 10 compared to sections exposed away from the dome, and concluded that most of the thinning is tectonic. As a result of such intense deformation the Chester core-rocks (see Appendix 4) exhibit a strong secondary banding parallel to the dome margins, yet there are localities within the dome where traces of the older layering are readily observed. Such isotopic and geologic traces of an older ancestry have not been observed in the Holts Ledge Gneiss in the Mascoma Dome. None of the episodes of metamorphism and deformation which affected the area subsequent to the deposition of the Ammonoosuc Volcanics (the most intense metamorphism recognized in the study area is garnet- to staurolite-grade) appears to have been of sufficient intensity to remove such traces from a Precambrian rock.

The zircon data for sample MHL-1, together with the field relationships disprove the conclusion of Chapman (1939) and Hadley (1942) that the Holts Ledge Gneiss is Devonian. The resolution of the present isotopic data is not sufficient to determine the age of the gneiss relative to the Ammonoosuc Volcanics. The data are consistent with the hypothesis based on field relationships that the Holts Ledge Gneiss lies beneath the Ammonoosuc Volcanics, but do not provide a strong check on the theory.

The author suggests that the Holts Ledge Gneiss has a volcanic origin and that the Pb^{207}/Pb^{206} age of 450 ± 25 m.y. measured on zircons from sample MHL - 1 dates the volcanic episode which produced the rock. Hopson (1964, p. 35) has shown that albitized and zeolitized volcanic clastic rocks commonly have compositions similar to those of the Holts Ledge Gneiss and some of the Ammonoosuc felsite layers (see Figure 12), whereas most common sedimentary rocks have different compositions. A volcanic origin could readily account for the nature of the layering observed in the Holts Ledge Gneiss, and the biotite-rich schlieren observed in some layers could be the remains of altered glassy bombs. These are preliminary conclusions, and the author is engaged in further geologic and isotopic studies on the origin of the Holts Ledge Gneiss.

Age and Origin of the Mascoma Group

About one fourth of the core-rocks of the Mascoma Dome are assigned to the Mascoma Group. Field relationships suggest these rocks are intrusive, and the author believes that the isotopic ages reflect the time of intrusion. The rocks have been deformed and metamorphosed subsequent to emplacement, but the effects of these later disturbances do not include significant granitization or remobilization. Most of the following discussion is concerned with the granitic rocks as they are the most abundant and best-studied rocks within the Mascoma Group.

U-Pb ages probably reflect the time of crystallization of zircon in the granitic rocks. As discussed previously this time appears to be 450 ± 25 m.y. The Rb-Sr whole-rock ages indicate the granitic rocks consolidated at about 440 ± 30 m.y. The two types of age measurements agree within experimental error. As previously discussed, it is not possible to establish the ages within narrower limits with the present data. Rb-Sr whole-rock ages on granitic rocks of the Lebanon Dome agree with those on granitic rocks of the Mascoma Group, suggesting the two granitic units are correlative within experimental error. The Rb-Sr ages on aplite and some of the fine-grained granite samples appear slightly lower than ages measured on the medium-grained granitic rocks. As discussed previously it is likely that these rocks, occurring as small bodies embedded in a matrix of rocks having less radiogenic Sr, could have had their apparent ages lowered during metamorphism.

The geologic and isotopic data demonstrate that the granitic rocks of the Mascoma Group are older than the Lower Silurian Clough Formation and are therefore not Devonian as concluded by Billings (1937, p. 501; 1956, p. 123-125). The resolution of the present data is not sufficient to determine the age of the granitic rocks of the Mascoma Group relative to Ammonoosuc Volcanics.

Several lines of evidence argue against significant metasomatism, granitization, or remobilization of the Mascoma Group during the Acadian (Devonian) orogeny or at any other time after deposition of the Clough Formation. The principal evidence is the agreement of the Rb-Sr and U-Pb ages. Rubidium is geochemically associated with potassium. If there had been significant addition of potassium (granitization) during the Devonian to an older, less potassic rock, the zircons might retain the older age, but the Rb-Sr whole-rock ages would be reduced at least in proportion to the amount of rubidium added, and might be entirely "reset" at the time of the younger event. The structure of the Mascoma Dome and the scarcity of minor folds in the lower parts of the mantle indicate that the core-rocks were competent during the major period of metamorphism and deformation. The granitic rocks lack any veined, streaky, or migmatitic appearance suggestive of later remobilization, and all of the observed crosscutting relationships can be satisfactorily attributed to events which occurred prior to deposition of the Clough Formation. The fact that even thin aplite veins retain Rb-Sr whole-rock ages in the range 415 to 440 m.y. is a further argument against Devonian or younger remobilization. It seems likely that during a metamorphic event sufficiently intense to remobilize the granitic rocks there would be sufficient exchange of Sr to "reset" the ages of aplite veins only a foot or so thick.

Some of these characteristics could be produced by intense in-place metamorphism of rhyolitic volcanic rocks, but the evidence suggests this has not occurred in the Mascoma Dome. Within the Holts Ledge Gneiss at least one felsite layer has the composition of rhyolite. If the main mass of granitic rocks had originated through the metamorphism of rhyolite, one would expect that this nearby layer would also become granitic in texture and appearance.

The geologic and isotopic data are consistent with a hypothesis that the granitic rocks consolidated from magma sometime within the interval 420 to 470 m.y. The igneous composition (see Figure 12) and appearance of the rocks; their massive, homogeneous, unstratified character; and the sharp, crosscutting contacts of the bodies in which they occur suggest the rocks are intrusive. Locally the quartz monzonite contains ovoid, biotite-rich schlieren which may be xenoliths of a more mafic rock. The crosshatch twinning of the microcline crystals previously described indicates that the potassium feldspar crystallized in an initially triclinic state (Laves, 1950, p. 150) probably indicating crystallization temperatures in excess of 500°C. The author suggests that the microcline crystals may have originated as phenocrysts and that their unusual exsolution features may be due to conditioning during metamorphism at temperatures near the monoclinic-triclinic inversion temperature. Figure 3 shows what appears to be a sill of quartz monzonite extending a short distance into the Holts Ledge Gneiss on central Moose Mountain. Otherwise, except for possible sills at localities 5 and 6 (Figure 3), the author saw no sills or dikes of the granitic rocks. Aplite veins are locally conspicuous in the granitic rocks, but nowhere were observed to crosscut other units.

A Volcanic and Plutonic Belt

The Oliverian Dome belt is part of a much larger belt characterized by major pre-mid Silurian volcanic and plutonic activity. It has long been recognized that the Ammonoosuc Volcanics can be traced most of the 250 mile length of the Oliverian belt, and more recently it has become apparent that they are in a broad sense correlative with similar volcanic rocks exposed along-strike as far away as extreme-northeastern Maine. Plutonic rocks ranging from gabbro to granite occur in the northern part of this belt (Pavlides and others, 1964, p. C 35; Albee, 1961, p. C 51), and are similar to the rocks of the Highlandcroft Series (Billings, 1937, p. 499-500; 1956, p. 46-48) exposed on the northwestern flank of the northern part of the Oliverian belt. The author's data indicate that the Mascoma Group, the Holts Ledge Gneiss, and probably correlative rocks in the other Oliverian Domes are also part of this complex. This greatly extends the magnitude of recognized pre-mid Silurian volcanic and plutonic activity in the southern part of the belt, as these rocks were previously considered Devonian. The volcanic belt lies within a geosyncline which is filled with a great thickness of early- and mid-Paleozoic rocks. Volcanic rocks are widely but sporadically distributed in space and time throughout the geosynclinal section but are best developed (in terms of both the volume of rock deposited and the persistence of volcanic activity through time) in a narrow belt extending from Long Island Sound through far-northeastern Maine. Pre-mid-Silurian plutonic activity is similarly maximized along this belt.

Future attention should be given to more detailed lithologic and chronologic subdivision of this complex. There may be a considerable range in ages among the rocks involved. Throughout the belt the volcanic and plutonic rocks are overlain by mostly non-volcanic strata ranging in age from late Lower Silurian to Lower Devonian, and in far-northeastern Maine Upper Ordovician (Ashgillian) non-volcanic strata (Boucot and others, 1964, p. 20-23) overlie the volcanic rocks. These relationahips indicate the minimum age of the main volcanic and plutonic activity.

Fossils ranging in age from late Early Ordovician (Arenigian; Neuman, 1964, p. E2) to late Middle Ordovician (Caradocian; Boucot and others, 1964, p. 19-20) have been identified in tuffs and in sedimentary rocks associated with the volcanic rocks. Radiometric ages of 418 $\pm$ 15 (Brookins and Hurley, 1965, p. 13) and 440 $\pm$ 30 (this report) have been measured on the Ammonoosuc Volcanics near Littleton, New Hampshire. Based on data presented earlier in this report, ages measured on some of the granitic plutons lie in a range 430 to 460 $\pm$ 20 to 40 m.y.

The origin of the magmas from which the granitic rocks were supposedly derived and their relations to the volcanic magmas are not well understood. The observation that the plutonic and volcanic rocks appear to be spatially and temporarily related suggests the possibility of genetic relationship. The granitic rocks constitute only a minor part of the total volume of igneous rock, mafic and intermediate rock being far more abundant. All of the rocks are possibly derived directly from mantle material. This is neither substantiated nor disproved by the present isotopic data, although the chemical data in Figure 12 suggest that the Holts Ledge Gneiss and the Mascoma Group are not related by "simple" differentiation. The alternative is to derive some or all of the rocks by fusion of crustal materials. The present Rb-Sr and U-Pb data reveal no traces of geological history prior to 500 m.y. in any of the rocks studied from the complex. Fusion of old, potassium-rich crustal materials with retention of the previously-generated radiogenic Sr would lead to raised values of the Sr^{87}/Sr^{86} ratio of initial Sr incorporated in the derived rocks.

The observed initial ratios, averaging 0.705 to 0.707, are not unusually high. This suggests, but does not prove, that the magmas were not derived from billion-year-old potassium-rich crustal rocks. Much more information is needed before the origin and relationships of the plutonic and volcanic rocks can be better understood. It is particularly necessary to learn more about the spatial and temporal relationships of the rocks, and about the metamorphic conditions which prevailed at the time of the plutonism. The author is engaged in further geologic and isotopic study of these problems.

Metamorphism in the Oliverian Belt

The major period of metamorphism recognized in the dome belt is post-Lower Devonian as it affected mantle-units of Lower Devonian age. In the part of the Mascoma Dome which was studied, the most intense metamorphism was garnet-to staurolite-grade. The metamorphic grade increases southward along the dome belt to where sillimanite or kyanite are common in pelitic rocks (Robinson, 1963). An upper limit on the age of metamorphism is provided by low-grade Triassic beds in the Connecticut Valley which contain clasts of metamorphosed Paleozoic rocks. Most geologists believe that the most intense metamorphism was Devonian, but many of the arguments that formerly supported this inference have been disproved. In northern Maine Boucot and others (1964, p. 73-74) have shown that the most intense metamorphism and deformation (chlorite-grade and upright, steep folding of strata) occurred between late Lower Devonian and late Middle Devonian time, but it is difficult to say if this relationship is true for the entire region.

The Rb-Sr mineral data on the Mascoma Dome suggest that there was an "event" about 250 million years ago associated with the redistribution of Sr among mineral gains in the rock samples. This effect in southern New England was first discussed by Faul and others (1963) and has also been noted by D. G. Brookins (oral communication) and Zartman, Snyder, Stern, Marvin, and Bucknam (1965), but the nature of the "event" has not yet been established. Wasserburg and Hayden (1955) and Brookins and Hurley (1965) have established that large pegmatites were emplaced in central Connecticut about 260 million years ago, and in the Narraganset Basin there is evidence for post-Pennsylvanian metamorphism (Quinn, 1953, p. 264). In central New England it has not yet been possible to attribute any major structural or metamorphic features to the "250 million year event." It is known that the disturbance of Rb-Sr mineral isochrons is a very sensitive indicator of metamorphism (Lanphere and others, 1964; Grant, 1964). One possibility is that there was a discrete thermal pulse about 250 million years ago that is recorded only in the disturbed geochronologic data. Alternatively the measurements may reflect the uplift and cooling of deep-seated rocks to a point where the isotopes of Sr lost their mobility.

The mineral data retain no trace of a major metamorphic event occurring in the Oliverian belt during the Devonian, but as indicated in the previous discussion of Rb-Sr mineral isochrons, such evidence is easily destroyed by "overprinting" of subsequent events. The author is presently engaged in Rb-Sr whole-rock and U-Pb zircon dating of plutonic rocks whose sequence in the metamorphic and depositional "program" has been established by field relationships. It is hoped that this work will provide more information on the sequence of events in the dome belt.

The Mantled Gneiss Dome Problem

Mantled gneiss domes appear to originate through middle-and high-grade metamorphism of rock sequences in which massive quartzo-feldspathic rocks are overlain by less competent, well-stratified rocks. There are potentially many geological situations which can establish such rock sequences, and the author's work indicates that mantled gneiss domes indeed have multiple origins. Some authors have felt that if a structure can be identified as a mantled gneiss dome there is strong presumptive evidence that the core-rocks are much older than the mantling strata and have been remobilized or granitized during later metamorphism. The work in New England indicates that such presumption is not universally valid.

Mantled gneiss domes are common in many areas where the deep-seated parts of former orogenic belts are exposed. The domes typically have cores of massive granite or gneiss encircled by concordant mantles of stratified metamorphic rocks. The massive core rocks contrast sharply with the stratified mantle rocks, and the origin of the core rocks is commonly a subject of considerable debate. The problem of mantled gneiss domes has its roots in difficulties which arise when geologists apply to the domes the criteria which are normally used to ascertain the relative ages of rock units, and becomes particularly acute when the core rocks are of igneous composition. The concordance of the core-mantle contacts suggests that the core rocks are older than the mantling strata, but local crosscutting relationships taken together with the igneous composition of the core rocks lead to precisely the opposite conclusion. Based chiefly on field study of such structures in Finland, Eskola (1949) proposed a theory of mantled gneiss dome formation which sought to reconcile the apparent dilemma posed by the field relationships. He summarized his theory in the following statement (Eskola, 1949, p. 461): "The mantled gneiss domes apparently represent earlier granite intrusions related to an orogenic period. The plutonic mass was later eroded and levelled, and thereafter followed a period of sedimentation. During a subsequent orogenic cycle the pluton was mobilized anew and new granitic magma was injected into the plutonic rock at the same time it was deformed into gneiss, causing its migmitization and granitization, or palingenesis." Eskola proposed that the following features were universal characteristics of mantled gneiss domes: (1) the domes mark sites which have been affected by two orogenic revolutions, (2) the contact between core and mantle is an unconformity, and (3) where cross cutting relationships are present they are the result of remobilization of the older rocks. He went further to suggest that the mantled gneiss dome mechanism provided a means by which older basement rocks might be transformed during later orogeny into young-looking plutonic bodies. The tectonic implications of the theory, the difficulty of determining the history of some of the individual domes, and the fact that some of the examples he cited had previously been interpreted in quite different ways contribute to the controversy surrounding Eskola's proposals. His formulation of the problem has stimulated much new thought and research, but it has never been clear that the validity of his theory has been tested sufficiently to justify applying his conclusions universally.

There is conclusive field and geochronologic evidence that many gneiss domes have formed through the action of later orogeny on older basement complex overlain unconformably by stratified rocks. Well studied examples are the Kuopio and Joensuu domes of Finland (Eskola, 1949; Preston, 1954; Wetherill, Kouvo, Tilton, and Gast, 1962; Kouvo and Tilton, 1966), the Baltimore Domes of Maryland (Broedel, 1937; Tilton and others, 1958; Hopson, 1964; Wetherill, Davis, and Lee-Hu, 1966), the World Beater Complex of California (Lanphere and others, 1963), and the Chester Dome group of Vermont (Thompson, 1950; Rosenfeld, 1954; Skehan, 1961; Faul and others, 1963; and the present work). The Vermont domes (see Appendix 4) are very good examples of this type of mantled gneiss dome, and the mechanism of progressive obliteration of the angularity of the core-mantle unconformity during mantled gneiss dome formation is particularly well displayed. The cores of many of these domes contain paragneisses as well as orthogneisses. This possibility was not anticipated in Eskola's original formulation of the dome problem, but few of his conclusions are altered if the theory is generalized to account for this. One misconception has arisen partly as a result of this generalization, and that is the concept that a long interval of time must necessarily elapse between the emplacement of the core rocks and their burial beneath the mantling strata. This is true for many of the examples studied but is not a necessary part of the mantled gneiss dome theory, and as pointed out by Wetherill (1965, oral communication) was not a part of Eskola's thinking on the subject.

It is interesting to note that very few geologists have ever suggested that the principal core-rocks of these domes are intrusive into the mantling strata. In all of these domes the core-rocks somehow "look" older, but before the advent of modern field and isotopic techniques it was difficult to give conclusive proof of this older age, and many of the arguments formerly advanced have proved to be specious. The Maryland and Vermont domes illustrate that deformation accompanying dome formation can almost completely obliterate the angular relationships which once existed at the unconformities separating the core- and mantle-rocks. In the Chester Dome the precise position of the unconformity is still undetermined, yet as described in Appendix 4 there can be little doubt that a major unconformity exists somewhere in the section, and its location is approximately known. Despite very intense deformation most of the core-rocks of these domes retain much trace evidence suggestive of a more ancient history. In both the Maryland and Vermont domes there are rocks which locally crosscut the mantling strata, but these rocks also crosscut stratification in the core-rocks, and can be distinguished from the primary core-rocks by differences in texture and mineralogy. Hopson (1964, p. 45-48) suggests that rocks of this type in the Maryland Domes (Gunpowder Granite) formed by partial fusion of the Baltimore Gneiss *in-situ*, whereas the present author feels that crosscutting rocks of this type in the Chester Dome (Baltimore

Trondhjemite) were intruded from depths below the level of present exposure. There is no proof that the trondhjemite formed by partial fusion of the Chester core-gneiss, although it remains a possibility. However these rocks formed, the fact remains that they are readily distinguishable from the primary core-rocks and their crosscutting relationships lead few geologists to suggest that the primary core-rocks are intrusive.

There exist other mantled gneiss domes whose origins have been more controversial. Their core-rocks are more massive and are igneous in composition and appearance. The core-rocks do not have the old "look" characteristic of the domes previously discussed. Local crosscutting relationships readily suggest that the entire core may be intrusive, and the domes are commonly so interpreted. Eskola (1949) suggested that these domes also contain older core-rocks, but argued that they are so intensely deformed that all traces of the older ancestry have been obliterated. The argument for the presence of older rocks is based on a plausible analogy, but is very hard to prove, as it contains a built-in "excuse" for not being able to demonstrate the proof. Previous geochronologic work on domes of this type has not fully settled the issue. Krough (1964) and Brookins and Hurley (1965) studied domes of this sort and found no isotopic evidence that the core-rocks were older than the mantling strata. Krough discussed very carefully and explicitly the possibilities that the Rb-Sr whole rock ages of the Burleigh Dome (Ontario) might have been reset during Grenville metamorphism. He concluded that the gneiss was probably not significantly older than the mantle-strata, but was unable to determine whether the core-rocks were slightly older or slightly younger than the mantle or whether there were crosscutting relations due to remobilization. Lacking detailed field relationships or a crosscheck by independent isotopic methods Krough was unable to rule out the possibility of massive Sr exchange during Grenville metamorphism and some possibility remained that Rb-Sr evidence for an older age of the core-rocks had been erased during very intense metamorphism.

The Oliverian Domes seemed well suited for a further study of this problem. A number of geologists had disagreed with the prevailing hypothesis that the core-rocks were Devonian intrusive rocks, and had argued that the core rocks might instead be Precambrian. The Rb-Sr study of Brookins and Hurley (1965) produced no evidence for the presence of old rocks in the cores of the Oliverian Domes in Connecticut. Their work did not resolve the ambiguities in the field relations which had led the diversity of opinion on the domes, and the possibility still remained that the dome-forming deformation and metamorphism might have been so severe as to make Precambrian core-rocks look much younger. The suitability of the Oliverian Domes, and the Mascoma Dome in particular, was further enhanced by the fact that they were singled out by Eskola (1949, p. 470-471) as specific examples of domes where the crosscutting relationships were caused by remobilization during the dome-forming metamorphism.

In the preceding parts of this paper the author demonstrated that the core-rocks of the Mascoma Dome are not Precambrian rocks disturbed at the time the domes were formed. It is further possible to show that the significant crosscutting relationships are the result of events which occurred before most of the mantle-units were deposited, and are therefore not due to remobilization at the time of formation of the domes. The observation that the core-rocks of the Oliverian Domes are not genetically related to those in the nearby Chester Dome demonstrates the uncertainty inherent in the reasoning-by-analogy on which Eskola's mantled gneiss dome hypothesis was based.

Recent studies of mantled gneiss domes indicate that the doming mechanism is not as effective in making "old rocks look young" as Eskola proposed. Modern field and isotopic techniques have given clear-cut proof of the old age of core-rocks in some domes which have been affected by kyanite-staurolite grade metamorphism and very intense deformation. The "signals" which indicate the age of these core-rocks are so strong that they should survive even significantly higher grades of metamorphism. In the Oliverian Domes plausible alternatives have been found to explain relationships which Eskola attributed to remobilization of older rocks at the time of dome formation. It is apparent that Eskola's arguments are not unique and that it is possible to have mantled gneiss domes in which none of the crosscutting relationships are due to remobilization during dome formation and in which the core-mantle contact is not an unconformity. The mantled gneiss dome theory provides few reliable clues to the history of dome rocks and each dome or group of domes must be studied individually. The term "mantled gneiss dome" should be retained as it is an elegant description, and will serve to remind the geologist studying such structures to test alternative hypotheses which might otherwise not enter his thinking.

APPENDIX 1

ANALYTICAL TECHNIQUES

Rb-Sr procedure

Sample treatment

Whole-rock samples (generally greater than 3kg in mass) were broken on a steel plate, crushed in a jaw crusher, and pulverized in a disk-grinder. The preparation room and equipment were carefully cleaned prior to running each sample. The crushed rock powder (finer than 50 mesh) was split down using sample splitters to yield at least four representative whole-rock samples which were stored in plastic vials. Mica samples were purified on the Franz magnetic separator, and other minerals were concentrated using both the Franz-separator and heavy liquid separation. Rather large (about 4 g.)whole-rock samples were dissolved to reduce the effects of sample inhomogeneity. Smaller amounts of mineral samples were dissolved -- the amount being calculated to yield 5-10 micrograms of Sr and Rb to load on the mass spectometer filament.

The chemical procedure is essentially that described by Wetherill, Tilton, Davis, and Aldrich (1956) and Zartman (1964). Initial solution was in Pt-ware and all subsequent treatment was in teflon-ware except for the use of pyrex-walled ion exchange columns and pyrex loading pipettes (see Wasserburg, Wen, and Aronson, 1964). The purified Rb and Sr solutions were loaded on outgassed Ta filaments and analyzed in a 12 inch 60 degree-sector Nier-type mass spectrometer previously described by Chow and McKinney (1956). The mass spectrometer is equipped with a multiplier. A 4 gram sample typically required the following quantities of reagents: 20 ml. 48-percent hydrofluoric acid (Baker Reagent grade); 6 ml. vycor distilled perchloric acid (G. F. Smith Co.); 200 ml. triple distilled water; and 100 ml. 2.5 N hydrochloric acid (prepared by passing HCl gas through triple distilled water). The greatest blanks measured were 0.037 micrograms Rb and 0.014 micrograms Sr in reagents sufficient for a 4 gram sample.

Concentration data and tracer calibrations

The Sr^{84}-enriched tracer used for all of the Sr analyses has the following isotopic composition, expressed in atom fractions: Sr^{84}(0.8344), Sr^{86}(0.0386), Sr^{87}(0.0132), and Sr^{88}(0.1138). Two tracer solutions were prepared from this material: a dilute tracer (Sr^{84}= 0.265 x 10^{-8} moles/gram) used primarily for aliquotted samples and analyses of low-Sr minerals, and a concentrated tracer (Sr^{84}= 17.92 x 10^{-8} moles/gram) used primarily for total-spiking whole-rock samples. Both tracers were calibrated by isotope dilution against gravimetrically prepared shelf solutions, and for both tracers the calibrations agree to within 0.4 percent with the gravimetrically estimated concentration of the tracer. The absolute concentration of the Sr tracer is probably known to within 0.6 percent. The Rb tracer (Rb 85/Rb 87 = 0.0084) was also prepared in dilute (Rb^{87}= 3.26 x 10^{-8} moles/gram) and concentrated (Rb^{87}= 78.58 x 10^{-8}moles/gram) form. The calibration of the Rb tracer is less accurate due to inability to correct for mass discrimination in the calibration runs. The average of calibrated concentrations of the Rb tracers agrees with the gravimetric estimate to within 0.3 percent but individual calibrations differed by as much as 0.8 percent. The absolute calibrations of the

Rb tracers are probably accurate to about 1 percent. No drift in the concentration of any of the tracers was detected during the course of the study.

The Rb/Sr ratios of replicated samples (Table 2) agree to within 2 to 3 percent, but since the ages on the same samples agree within closer limits (0.5 to 2 percent) some of the apparent error in reproducibility may be due to sample inhomogeneity. Error brackets on the figures are plotted to show the effects of 4 percent uncertainty in the Rb/Sr ratios.

Measurement of Sr^{87}/Sr^{86}

Sr^{87}/Sr^{86} composition of most samples was determined by calculation from data on spiked runs according to equations given in Appendix 3. The Sr composition of some of the earlier-run samples was measured directly on unspiked aliquots of the sample solution, but this procedure has the disadvantage of requiring that chemical yields of Sr and Rb be 100 percent in all stages of handling the sample prior to mixing in the tracer. Determinations by both of these methods agree to within less than 0.3 percent and replicate determinations by a single method also agree within this range. The absolute accuracy of this measurement is not critical to this study but is believed to be within 0.3 percent. All Sr^{87}/Sr^{86} values are normalized to $Sr^{86}/Sr^{88} = 0.1194$, using the iterative procedure outlined in Appendix 3.

Zircon Procedure

Zircons were concentrated from 50 to 150 kg crushed-rock samples using the Wilfley Table, Franz separator, and heavy liquid procedures. Prior to fusion the concentrates were washed successively (20 minutes each) in steaming, reagent-grade HNO_3, HCl, and double-distilled HNO_3. The chemical extraction and sample loading procedures closely follow those described by Tilton, Davis, Wetherill, and Aldrich (1957). The dissolved sample was split into three aliquots. One of these aliquots was spiked for Pb, one for U, and one was left un-spiked for determination of the isotopic composition of Pb. The analyses were made on the CIT #1 12" mass spectrometer previously described, except for the composition-run on sample MMQ - 11 - Zr - mag which was made on the CIT #2 12" machine. The Pb^{207}/Pb^{206} ratio of this sample was determined three times with the following results: 0.05579 (spiked-run on CIT #1 M.S. with electron multiplier), 0.05564 (composition-run on CIT #2 M.S. with electron multiplier), and 0.05551 (composition-run on CIT #2 M.S. with simple collector). The total variation in these measurements is one half of the $\pm$ 0.5 percent error in this ratio indicated in Table 3 and Figure 11. The uranium blank is about 0.01 micrograms per analysis (R.H. Steiger, oral communication, 1967). The Pb^{204}/Pb^{206} ratios (Table 3) are sufficiently low that uncertainties in the composition of blank plus common Pb result in only second-order errors. Sufficient Pb has not yet been extracted from the reagents used to permit determination of the isotopic composition of contamination Pb. The analyses were corrected by attributing all of the Pb^{204} to common Pb of the composition indicated in Table 3 based on determinations of mid-Paleozoic common Pb in the central Appalachians by Doe and others (1965, p. 1959). Appropriate scale-factor and square root of the mass-ratio corrections were applied to the measured ratios. At the time of this writing the author's experiments to determine the precision and accuracy of the U and Pb concentrations are incomplete. Error limits are not quoted for these values, and the interpretation of the ages was not based heavily on the concentration data.

APPENDIX 2

SAMPLE LOCALITIES

The samples analyzed in this study have been assigned sample codes of the form: LCR - 5 - w_{II}. The first three letters are a code referring to the sample locality. The first letter indicates the following:

G	Green Mountain Anticlinorium
C	Chester Dome
L	Lebanon Dome
M	Mascoma Dome
A	Ammonoosuc Volcanics
H,F	Highlandcroft Series.

Each rock collected from a locality is assigned an arabic number. The final lower case letter indicates the character of the sample (w= whole-rock sample; other letters indicate minerals as explained in Table 2.) Subscript Roman numerals are assigned to separate splits of the sample. In the sample descriptions the major minerals are listed in order of decreasing abundance, and the state of alteration of the rock is briefly characterized. Coordinates are given in degrees and minutes, followed by the name of the 15 minute quadrangle in which the locality occurs.

Locality 1. (43°16.29'N, 72°32.78'W; Ludlow) Locality shows reversed-sense drag folds in core gneiss of Chester Dome.

Locality 2. See locality CSG.

Locality 3. (43°20.86'N, 72°35.20'W; Ludlow) This locality is a large stock of trondhjemite containing rotated xenoliths of the

Gassetts Schist and other mantle units.

GBF. ($43^{o}26.11'$N, $72^{o}43.89'$W; Ludlow) Interlayered dark schist and metasedimentary quartzite cut by veins of pegmatite are exposed in a roadcut on U.S. Route 103 immediately west of the bridge over Branch Brook at Buttermilk Falls. All of the rocks are fresh.

No. 1. (11 kg.) Coarsely granulated plagiocalse-quartz-microcline-muscovite pegmatite crosscuts metasedimentary quartzite.

Nos. 7 (5 kg), 8 (14 kg), 9 (15 kg), and 10 (8 kg). The most abundant rock in the roadcut is dark, tourmaline bearing, graphitic muscovite-quartz-plagioclase schist.

GDH. ($43^{o}29.01'$N, $72^{o}42.92'$W; Ludlow) At the summit of dry hill occurs a pluton of fine-grained granite. The granite is sheared and jointed, and truly fresh pieces could not be obtained.

Nos. 2 (2 kg), 3(3 kg), and 4 (25 kg.). The granites consist of quartz, microcline, and sericitized plagioclase with minor amounts of biotite and calcite.

GMH. ($43^{o}22.06'$N, $72^{o}47.48'$W; Wallingford) A large roadcut 0.1 miles north of the Rutland Co. line on Vt. Route 155 exposes interlayered light and dark heterogeneous gneiss (Mt. Holly Complex). The roadcut is less than 10 years old and the rocks appear fresh.

No. 3 (52 kg). This rock is light-colored, fine-grained, felsitic plagioclase-quartz gneiss with minor quantities of biotite and muscovite.

No. 4 (23 kg). Interlayered with the light-colored gneiss is a band of dark, fine-grained plagioclase-quartz biotite gneiss with minor amounts of muscovite.

CSG. ($43^{\circ}19.07'$N, $72^{\circ}36.52'$W; Ludlow) Heterogeneous gneiss in the core of the Chester Dome is exposed in a long roadcut on U. S. Route 103 about one half mile SE of Gassetts, Vermont. The roadcut was made in 1960-61 and all of the rocks appear fresh. All of the minerals are slightly elongated parallel to foliation.

Nos. 1 (11 kg), and 2 (7 kg). The south end of the outcrop consists of interlayered fine grained light-colored (plagioclase-quartz-biotite-epidote) and dark colored (plagioclase-epidote-quartz-biotite) gneiss.

No. 4 (3 kg). The most common rock is massive, medium-grained, plagioclase-quartz-muscovite-epidote-biotite gneiss locally containing traces of microcline.

No. 5 (30 kg) Occasional shallow-dipping layers of dark (epidote-biotite-quartz) gneiss occur near the north end of the roadcut.

No. 6 (3 kg). The massive gneiss is cut by a 12 foot thick dike of leucocratic quartz diorite (plagioclase quartz-muscovite).

LBR. ($43^{\circ}40.43'$N, $72^{\circ}14.23'$W; Mascoma). No. 1 (18 kg). Border gneiss of the Lebanon Dome is exposed in a small roadcut on the east side of the Etna Road 0.2 mi. south of the Hanover town line. The rock consists of intensely sericitized plagioclase, quartz, microcline, biotite, and muscovite.

LCV. ($43^{\circ}42.03'$N, $72^{\circ}16.48'$W; Hanover) No. 1. (20 kg.) Typical medium-grained granite of the granitic sub-core of the Lebanon Dome is exposed in a small, recent roadcut on the east side of Rayton Road in Hanover, N.H. The rock appears fresh although the plagioclase is slightly sericitized. Microcline occurs in somewhat elongate 1 cm. crystals set in a matrix of medium-grained plagioclase and quartz. Muscovite occurs in oval shaped aggregates of well foliated flakes. These features are typical of the other granites in the central core of the Lebanon Dome.

LCQ. (43°39.12'N, 72°15.23'W; Hanover) No. 1 (20 kg). Typical Lebanon core granite is exposed in an abandoned quarry (el. 600') on the east side of Quarry Hill in Lebanon, N.H. The rock appears fresh but the plagioclase is slightly sericitized.

LCW. (43°38.51'N, 72°16.09'W; Hanover) No. 1 (15 kg). A roadcut on Interstate 89 west of Lebanon, N.H., exposes fine-grained, leucocratic granite which J. B. Lyons (personal communication, 1965) reports is typical of rocks occurring along the southeast margin of the central core of the Lebanon Dome. The roadcut was constructed during 1965-66 and the rock is fresh.

LCR. (43°39.91'N, 72°14.75'W; Mascoma) A large roadcut on relocated N.H. Route 120 north of Lebanon, N.H., exposes typical Lebanon core granite. The cut was blasted in 1964-65 and the rock is fresh. The following samples were collected from blocks blasted during the construction.

Nos. 1 (5 kg) and 3(17 kg) are samples of leucocratic aplite which vein the more typical granite.

No. 5 (29 kg) is a sample of medium-grained granite representative of most of the material in the cut. The plagioclase is slightly sericitized.

MWL. (43°46.30'N, 72°5.07'W; Mt. Cube) No. 1 (11 kg). Ledges (el. 1500') on the north side of trail extending ESE from x1233 expose massive, fine-grained quartz-plagioclase-biotite gneiss typical of the Holts Ledge Gneiss. The plagioclase is sericitized.

MHL. (43°46.36'N, 72°5.75'W; Mt. Cube) No. 1 (99 kg). More gneiss typical of the Holts Ledge Gneiss is exposed by low cliff ledges along a woods road 1.8 mi. S. of fork in road (el. 880') near Dartmouth Skiway, Lyme Center, N.H. The rock is massive, faintly laminated, very fresh looking plagioclase-quartz-biotite gneiss. Zircons separated from the rock do not comprise a uniform suite, but do not appear to be a mixture of two or more distinct populations. The zircons are colorless, somewhat irregular, faceted crystals with locally cloudy opaque patches in their interiors.

MMS (43°39.90'N, 72°8.67'W; Mascoma) No. 2. (12 kg) An abandoned quarry at the end of May Street north of Enfield, N.H., exposes massive quartz monzonite typical of the Mascoma Group. Microcline occurs in 1 cm somewhat irregular crystals which display the unusual type of exsolution described on page 18. The plagioclase and quartz occur in a medium-grained matrix, and the rock also contains strongly foliated biotite and well developed 1-2 mm octahedra of magnetite. Myrmekitic intergrowth of feldspar and quartz is common. The rock is fresh.

MMQ (43°41.16'N, 72°9.11'W; Mascoma). More massive quartz monzonite typical of the Mascoma Group is exposed in an abandoned granite quarry.

No. 11 (60 kg) is a sample of the massive quartz monzonite similar to MMS-2, above. Zircons from this rock are colorless, slightly irregular, faceted crystals commonly with cloudy opaque patches in the interior.

No. 3 (2 kg) is a sample from an aplite vein which cuts the massive quartz monzonite. Both rocks appear fresh, but the plagioclase in MMQ-11 is slightly sericitized.

MJH (43°36.92'N, 72°6.14'W; Mascoma) No. 1 (60 kg). A small roadcut 0.3 miles east of the crossroads SW of Jones Hill exposes fine-grained quartz monzonite consisting of microcline, quartz, plagioclase, and biotite. Zircons separated from this rock are similar to those from MMQ-11 except that the opaque cloudy patches are fewer and smaller. Both suites of zircons are heterogeneous, but neither seems to be a mixture of two or more discrete populations.

MCP. (43°34.48'N, 72°6.92'W; Mascoma) No. 1 (14 kg) Fine-grained granite is exposed in a stream west of George Pond. The granite looks fairly fresh but is somewhat granulated and occurs near a Triassic(?) fault. Major constituents are slightly sericitized plagioclase, quartz, and microcline.

HHF. (44°19'N, 71°50'W; Littleton) No. 1 (18 kg). A road-cut on re-routed N.H. Route 18 exposes brecciated quartz monzonite of the Highlandcroft Series. The rock is similar to sample FSM-1 described below, except that the plagioclase in HHF-1 is intensely sericitized.

FSM (43°56.11'N, 72°7.22'W; Mt. Cube) A roadcut on U.S. Route 5 at the junction with a secondary road leading to the Sawyer Mtn. picnic area exposes typical samples of the Fairlee Quartz Monzonite. Nos. 1 (11 kg) and 4 (12 kg) are greenish-gray granitic rocks consisting of plagioclase, blue quartz, microcline, biotite, chlorite, and traces of calcite.

AWR.($44^{\circ}19.63'$N, $71^{\circ}51.79'$ W; Littleton) Nos. 1 (6 kg), 2 (7 kg), and 6 (2 kg) are well stratified greenstones of the Ammonoosuc Volcanics occurring in a roadcut at the junction of N.H. Route 18 and Williams Road west of Littleton, N.H. Principle minerals are fine-grained, sericitized plagioclase, quartz, and chlorite.

ASH.($44^{\circ}14.20'$N, $71^{\circ}53.00'$W; Moosilauke) Nos. 1 (7 kg), 2 (2 kg), and 3 (1 kg) are samples of fine-grained Ammonoosuc felsite from a roadcut on N.H. Route 10 at the junction with the Sugar Hill Road.

APPENDIX 3

ISOTOPE DILUTION EQUATIONS

Concentration and composition equations

Let A_i, B_i, C_i be the atomic abundances (in moles) of isotopes a,b,c of an element in substance i. Then $(\frac{A}{B})_i$ is the atomic ratio of nuclides a and b in that substance. The subscripts are used as follows: t for tracer (spike), s for sample, and m for mixture of tracer and sample. All of the equations which follow are derived from the relation:

$$\left(\frac{A}{B}\right)_m = \frac{As + At}{Bs + Bt} \,. \tag{1}$$

This may be solved two ways to give the concentration of the sample from the concentration of the tracer:

$$A_s = \left[\frac{(B/A)m - (B/A)t}{(B/A)s - (B/A)m}\right] A_t \qquad \text{or,} \tag{2}$$

$$A_s = \left[\frac{(A/B)m - (A/B)t}{1 - (A/B)m\ (B/A)s}\right] B_t \tag{3}$$

If equation (2) is also expressed using the ratios (C/A), then

$$\frac{At}{As} = \frac{(B/A)s - (B/A)m}{(B/A)m - (B/A)t} = \frac{(C/A)s-(C/A)m}{(C/A)m-(C/A)t} \tag{4}$$

Important relations of the form

$$(B/A)s = (B/A)m + \left(\frac{(B/A)m-(B/A)t \quad (C/A)s-(C/A)m}{(C/A)m - (C/A)t}\right) \tag{5}$$

are derived from equation (4).

Some important properties of these equations regarding the propagation of errors appear when numerical values are inserted. Each equation has terms of the form

$$\frac{(k)s - (x)m}{(x)m - (k)t} \quad , \tag{6}$$

where by choice of the proportions of tracer mixed with the sample $(x)_m$ may assume any value between $(k)_s$ and $(k)_t$. If two large numbers are subtracted to give a small number any uncertainties in the large numbers will be magnified in the difference. Errors of this sort grow without limit as $(x)_m$ approaches either $(k)_s$ or $(k)_t$. If uncertainties in each of the parameters are equivalent these errors are minimized by setting $x)_m = ((k)s(k)t)^{\frac{1}{2}}$ (Webster, 1960, p. 216).

Sometimes one of the parameters is measured with greater uncertainty than the others. Often it is possible to find a form of the equations and a value of $(x)_m$ which minimize the effect of uncertainties in a particular parameter, and conversely. The more favorable solutions are found by inserting numerical values for the ratios and inspecting alternate arrangements of the equations. Relations of the form of equation (5) are more reliable if the term on the far right is kept small compared to the $(\frac{B}{A})m$ term.

The ratios $(k)_t$ are hard to measure reliably if the tracer is composed almost entirely of a single isotope, and in such cases the equations should be chosen to minimize the effects of uncertainties in the tracer composition. Suppose the tracer is rich in b and the sample is rich in a. Equation (3),

$$A_s = \left[\frac{(A/B)m - (A/B)t}{1 - (A/B)m(B/A)s}\right] B_t$$

minimizes the effect of uncertainty in $(\frac{A}{B})t$ for suitable values of $(\frac{A}{B})m$. The uncertainty is magnified in the inverse equation

$$Bs = \left[\frac{(B/A)t - (B/A)m}{(B/A)m\ (A/B)s - 1}\right] At \qquad (7)$$

The situation becomes worse as the purity of b in the tracer is increased and equation (7) cannot be used at all if the tracer consists entirely of isotope b. Errors in Bs can be minimized if the tracer was calibrated using the reciprocal of equation (7) and $(\frac{B}{A})m$ in the sample run is held close to the value $(\frac{B}{A})m$ used in calibrating the tracer, but these troubles are avoided entirely if equation (3) is used instead. For this case equation (2) is bad for the same reasons as equation (7).

Discrimination

The equations in the preceeding section have as parameters ratios of the numbers of atoms in isotope pairs, but the quantities actually measured are ratios of intensities of mass spectrometer peaks corresponding to these isotopes. This section is concerned with the problem of converting from the measured intensity ratios to the needed ratios of atoms.

The intensity (I_i) of a peak measured in the mass spectrometer is related to the number of atoms (a_i) of nuclide i by $I_i = S_i \cdot a_i$, where S_i is an incompletely known function of many variables including the chemistry of the nuclide and the behavior of the mass spectrometer. Many of these variables cancel if measurements are restricted to ratios of isotopes of single elements, but even among isotopes of a single element, S is a mass dependent function S(m) and measured ratios of intensities differ from the actual ratios of numbers of atoms by a mass discrimination factor

$$\alpha\ (m_1,m_2) = \frac{a_{m1}/a_{m2}}{I_{m1}/I_{m2}} \quad . \qquad (8)$$

Without knowing very much about the function S it is possible to deduce useful information regarding the factor α. If Si is independent of the properties of other isotopes then an important relation holds:

$$\alpha\,(m_1,m_3) = \frac{S(m_3)}{S(m_3)} = \frac{S(m_2)}{S(m_1)} \cdot \frac{S(m_3)}{S(m_2)} = \alpha(m_1,m_2) \cdot \alpha(m_2,m_3) \qquad (9)$$

In a good instrument $\alpha\,(m_1,\ m_1 + \Delta\,m) = (1 + \delta)$ where δ is small compared to 1, and $\alpha\,(m_1 + \Delta\,m,\ m_1 + 2\Delta\,m)$ differs from $(1 + \delta)$ only by higher order terms which are neglected. In this case the relation

$$\alpha(m_1, m_1 + k \cdot \Delta m) = (1 + \delta)^k \qquad (10)$$

follows directly from equation (9).

In solid source mass spectrometry the value of δ generally varies from run to run and may fluctuate within a single run. These variations in δ can possibly be minimized by the use of a multiple filament ionization source under rigorously controlled conditions (Shields, Garner, Hedge, and Goldich, 1963). For precise work it is desirable to attempt a correction for mass discrimination, but each of the procedures for doing this can introduce secondary errors which are discussed below. It is necessary to evaluate these errors and to guard against conditions where these errors become so large that they outweigh the advantage of applying the correction.

Mass discrimination α may be calculated through the use of a double spiking procedure outlined below in its simplest form. Prepare a tracer rich in two isotopes which are rare in the sample and mix the sample with so much of this tracer that the ratio of these two isotopes in the mixture is essentially determined by their ratio in the tracer. Mass discrimination is calculated by comparing the measured and calibrated ratios of these two isotopes and a correction based on equation (10) is applied to the other isotope ratios.

Some elements have several isotopes (common isotopes) whose relative abundance is nearly constant in natural samples. (Common isotopes are stable or long lived nuclides having no long lived radioactive progenetors.) For these elements one may calculate the discrimination affecting a given spiked or unspiked run. For an unspiked sample the procedure is straight-forward. Compare the intensity ratio of a pair of common isotopes with the accepted value of their abundance ratio to calculate the mass discrimination factor (equation 8) and use equation (10) to correct the other ratios. It is essential to report both the accepted value (a_{m1}/a_{m2}) of the common isotope ratio used for this normalization and the procedure used for the normalization to facilitate interlaboratory comparisons: all workers should agree on a common accepted value for (a_{m1}/a_{m2}). A systematic error is introduced if (a_{m1}/a_{m2}) is incorrectly measured, but older data so affected can be readily corrected if a more accurate measurement of (a_{m1}/a_{m2}) becomes available. It is interesting to note that the effect of such a systematic error is entirely analogous to errors arising from uncertainties in the decay constants used in geochronology. In some cases a value of (a_{m1}/a_{m2}) enters the derivation of a particular decay constant and for internal consistency the same value of (a_{m1}/a_{m2}) should be used to calculate discrimination.

Another type of error is introduced if the abundance of the common isotopes varies in nature. Heavy isotopes (A greater than about 40) are probably not seriously fractionated by normal geologic processes. For example, Shields and others (1963) carefully measured (Rb^{85}/Rb^{87}) ratios in a wide variety of natural mineral samples of different ages and origins and found no variations outside experimental error ($\pm$ 0.0049, 95 percent confidence limit). For lighter elements it is necessary to resort to the double spiking procedure described previously. For heavier elements the author prefers to rely on the constancy of the natural abundances of common isotopes.

If an element has three common isotopes whose abundance is constant in natural samples the following values can be determined from a single mass spectrometer run on mixture of the sample plus a tracer enriched in a single isotope: (1) mass discrimination affecting the particular run, (2) concentration of the element in the sample, and (3) the abundance ratios of any radiogenically produced isotopes in the sample. The principle of the discrimination correction is identical to that described above and the same conclusions apply strictly. At this point a generalized nomenclature becomes cumbersome and the remainder of the treatment is derived for a particular example. The principles remain generally applicable.

The common isotopes of Sr are Sr^{84}, Sr^{86}, and Sr^{88}. From equation (4) it follows that

$$\frac{\left(\frac{Sr^{88}}{Sr^{84}}\right)_s - \left(\frac{Sr^{88}}{Sr^{84}}\right)_m}{\left(\frac{Sr^{88}}{Sr^{84}}\right)_m - \left(\frac{Sr^{88}}{Sr^{84}}\right)_t} = \frac{\left(\frac{Sr^{86}}{Sr^{84}}\right)_s - \left(\frac{Sr^{86}}{Sr^{84}}\right)_m}{\left(\frac{Sr^{86}}{Sr^{84}}\right)_m - \left(\frac{Sr^{86}}{Sr^{84}}\right)_t} \qquad (11)$$

where the symbols for the nuclides stand for the atomic abundances (a_i) in numbers of atoms and the subscripts are used as before. Due to mass discrimination the measured ratios of intensities for the mixture $(I_i/I_j)m$ differ from the atomic ratios $(a_i/a_j)m$ for which equation (11) is strictly true. These are related through by equations of the form

$$(a_i/a_j) = \alpha(i,j)\cdot(I_i/I_j) \qquad \text{(cf. eqtn. 8)}$$

From equations (8,10, and 11) we obtain

$$\frac{\left(\frac{Sr^{88}}{Sr^{84}}\right)_s - \left(\frac{Sr^{88}}{Sr^{84}}\right)_m^* \cdot \alpha^2}{\left(\frac{Sr^{88}}{Sr^{84}}\right)_m^* \cdot \alpha^2 - \left(\frac{Sr^{88}}{Sr^{84}}\right)_t} = \frac{\left(\frac{Sr^{86}}{Sr^{84}}\right)_s - \left(\frac{Sr^{86}}{Sr^{84}}\right)_m^* \cdot \alpha}{\left(\frac{Sr^{86}}{Sr^{84}}\right)_m^* \cdot \alpha \cdot \left(\frac{Sr^{86}}{Sr^{84}}\right)_t} \qquad (12)$$

where $\left(\frac{Sr^{88}}{Sr^{84}}\right)^*$ and $\left(\frac{Sr^{86}}{Sr^{84}}\right)^*$ are the measured ratios of intensities of ion beams; then

$$\alpha^2 \cdot A + \alpha \cdot B + C = 0$$

$$= \frac{-B \pm \sqrt{B^2 - 4\,AC}}{2A} \qquad (13)$$

where $A = \left(\frac{Sr^{88}}{Sr^{84}}\right)_m^* \cdot \left[\left(\frac{Sr^{86}}{Sr^{84}}\right)_t - \left(\frac{Sr^{86}}{Sr^{84}}\right)_s\right]$

$B = \left(\frac{Sr^{86}}{Sr^{84}}\right)_m^* \cdot \left[\left(\frac{Sr^{88}}{Sr^{84}}\right)_s - \left(\frac{Sr^{88}}{Sr^{84}}\right)_t\right]$

$C = \left(\frac{Sr^{86}}{Sr^{84}}\right)_s \cdot \left(\frac{Sr^{88}}{Sr^{84}}\right)_t - \left(\frac{Sr^{88}}{Sr^{84}}\right)_s \cdot \left(\frac{Sr^{86}}{Sr^{84}}\right)_t$.

While equation (13) shows the correct functional dependence of α, it is inconvenient to use directly, but there are several alternative approaches. Equation (13) may be solved by substituting α = $(1 + \delta)$ and neglecting higher order terms to give

$$\alpha_{2\ mass} = -\frac{A+B+C}{2A+B} \qquad (14)$$

Alternately discrimination may be solved by the following simple iterative procedure which is well suited for automated data processing.

Substituting into equation (5),

$$\left(\frac{Sr^{88}}{Sr^{86}}\right)_s^{calc} = \left(\frac{Sr^{88}}{Sr^{86}}\right)_m + \frac{\left[\left(\frac{Sr^{88}}{Sr^{86}}\right)_m - \left(\frac{Sr^{88}}{Sr^{86}}\right)_t\right]\left[\left(\frac{Sr^{84}}{Sr^{86}}\right)_m - \left(\frac{Sr^{84}}{Sr^{86}}\right)_s\right]}{\left(\frac{Sr^{84}}{Sr^{86}}\right)_t - \left(\frac{Sr^{84}}{Sr^{86}}\right)_m} \qquad (15)$$

The iteration is begun by assuming $\alpha = 1$ and solving equation (15).

Estimate α from the relation

$$\alpha_{2\ mass} = \frac{\left(\frac{Sr^{88}}{Sr^{86}}\right)_s^{accepted}}{\left(\frac{Sr^{88}}{Sr^{86}}\right)^{calculated}} \qquad (16)$$

and correct the measured intensity relations using equation (10).

The iteration may be continued until the changes in α become negligibly small.

The analysis of Rb.

Rb has only two isotopes suitable for mass spectrometric analysis (Rb^{8} , 72%; Rb^{87}, 28%). Measurement of the concentration of Rb^{87} in the sample is required for Rb-Sr dating. This analysis is most commonly done by isotope dilution using a tracer enriched in Rb^{87} and solving

$$(Rb^{87})s = \left(\frac{(Rb85/Rb87)m - (Rb85/Rb87)t}{(Rb85/Rb87)s - (Rb85/Rb87)m}\right)(Rb^{87})t \qquad (17)$$

Equation (17), derived from equation (2), gives the strongest solution for an Rb^{87} enriched tracer. Other possible solutions unfavorably magnify uncertainties in the ratio $(\frac{Rb^{85}}{Rb^{87}})_t$ which increase with increasing isotopic purity of the tracer. The reasons for this are given in the discussion following derivation of equations (1-5).

A slight theoretical advantage applies to using an Rb^{85} rich tracer and basing the solution on equation (3) since this procedure demagnifies uncertainties in the isotopic composition of both the sample and the tracer (see Crouch & Webster, 1963, p. 129). There may be no practical gain from the procedure though, because: (1) the optimum ratio $(\frac{Rb^{85}}{Rb^{87}})m$ is in a range more difficult to measure than is the case for equation (2) with Rb^{87}-rich tracer, and (2) according to Shields and others (1963) the natural variations in $(\frac{Rb^{85}}{Rb^{87}})s$ are smaller than the uncertainties in present day measurements of $(\frac{Rb^{85}}{Rb^{87}})m$ due to variable discrimination.

Because Rb has only two common isotopes there is no way of calculating the amount of mass discrimination affecting a given mass spectrometer run. An average discrimination factor may be computed by comparing the average of compositions of mineral Rb measured on a given instrument with an accepted value of $(\frac{Rb^{85}}{Rb^{87}})s$. For many machines it is advantageous to correct all measured Rb ratios by this average discrimination correction.

The value $(\frac{Rb^{85}}{Rb^{87}})s = 2.591$ (Nier, 1950, p. 451) was used in calculating both commonly used Rb decay constants, $= 1.39x10^{-11}$ year^{-1} (Aldrich, Wetherill, Tilton, and Davis, 1956) and $= 1.47x10^{-11}$ (Flynn and Glendenin, 1959). If the value $(\frac{Rb^{85}}{Rb^{87}})s = 2.5995 \pm 0.0015$ (Shields and others, 1963) is used both decay constants should be increased by 0.23 percent.

The analysis of Sr.

Strontium has three common isotopes, Sr^{84}, Sr^{86}, and Sr^{88}, and a fourth isotope, Sr^{87}, whose abundance varies due to production from the B-decay of Rb^{87}. The following quantities of interest in geochronological work may be determined from a single mass spectrometer run on a mixture of sample and tracers: 1) the proper correction for mass discrimination 2) the concentration of Sr^{86} in the sample, and 3) the ratio $(\frac{Sr^{87}}{Sr^{86}})$ in the sample. As shown below this can be accomplished best through the use of a tracer enriched in the single isotope Sr^{84}.

There is a signal advantage to making all the measurements on a single run, as this avoids the need for splitting the sample into aliquots for the separate determination of isotopic composition and concentration. It is difficult, particularly in the case of total rock samples, to obtain homogeneous splits of samples before they have been dissolved. Good splits may be obtained by aliquoting the sample in solution but the advantages of adding the tracer during dissolution are lost. The procedure outlined below has neither disadvantage.

The author recommends the following procedure for the determination of Sr. Before dissolving the sample add tracer enriched in Sr^{84} and Rb tracer, and prepare the sample for mass spectrometry. Correct the mass spectrometer data for fractionation using equation (15) or one of the alternate solutions. Use equation (3) in the following manner to solve for the concentration of Sr:

$$(Sr^{86})_s = \left[\frac{\left(\frac{Sr^{86}}{Sr^{84}}\right)_m - \left(\frac{Sr^{86}}{Sr^{84}}\right)_t}{1 - \left(\frac{Sr^{86}}{Sr^{84}}\right)_m \left(\frac{Sr^{84}}{Sr^{86}}\right)_s}\right] \cdot (Sr^{84})_t \qquad (18)$$

The ratio $\left(\frac{Sr^{87}}{Sr^{86}}\right)_s$ is found by substituting into equation (5):

$$\left(\frac{Sr^{87}}{Sr^{86}}\right)_s = \left(\frac{Sr^{88}}{Sr^{86}}\right)_s \left[\left(\frac{Sr^{87}}{Sr^{88}}\right)_m + \frac{\left(\left(\frac{Sr^{88}}{Sr^{86}}\right)_m - \left(\frac{Sr^{88}}{Sr^{86}}\right)_t\right)\left(\left(\frac{Sr^{84}}{Sr^{86}}\right)_s - \left(\frac{Sr^{84}}{Sr^{86}}\right)_m\right)}{\left(\frac{Sr^{84}}{Sr^{86}}\right)_m - \left(\frac{Sr^{84}}{Sr^{86}}\right)_t}\right] \qquad (19)$$

The sample and tracer should be mixed to give

$$\left(\frac{Sr^{84}}{Sr^{86}}\right) \cong 1.$$

APPENDIX 4

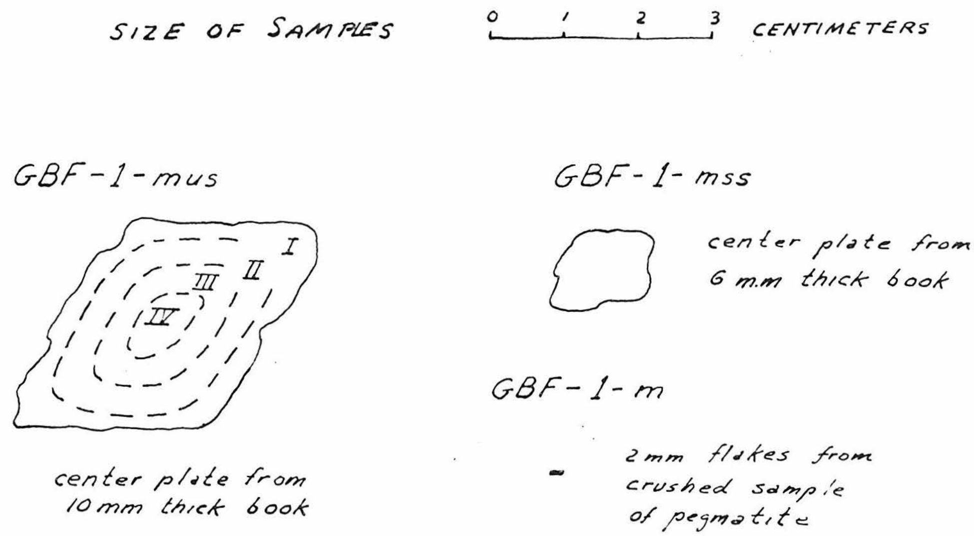
CHESTER DOME GROUP AND RELATED ROCKS

The Chester Dome is the largest of a group of mantled gneiss domes occurring in a north-trending belt a few miles west of the Connecticut River in central New England (Figure 1). The two northernmost domes are expressed chiefly by an arched pattern in the mantling strata, but in the Chester Dome and other domes further south massive core gneisses are exposed at the surface. Previous workers (Thompson, 1950; Doll and others, 1961) have shown that the core gneiss of the Chester Dome correlates with the Precambrian Mt. Holly Complex exposed to the west in the core of the Green Mountain Anticlinorium, which in turn appears equivalent to some of the units exposed still further west in the Adirondack Massif. The Precambrian age of the Mt. Holly Complex is readily deduced from field evidence (Thompson, 1950) and is confirmed by geochronologic data of Faul, Stern, Thomas, and Elmore (1963) and the present report. The Mt. Holly Complex consists of highly metamorphosed rocks overlain with profound structural and metamorphic unconformity by less strongly metamorphosed strata of later Precambrian(?), Lower Cambrian, and younger age. The Precambrian age of the core rocks of the Chester Dome is also indicated by field relations (Thompson, 1950). All but the lowest Paleozoic strata flanking the Green Mountain Anticlinorium have been traced directly to the flanks of the Chester Dome, and some of the lowest units which do not connect at the surface have been identified by matching the sections. The core gneisses are similar to gneisses in the core of the Green Mountains. The structural unconformity has been obliterated.

Obliteration of the unconformity between the core and mantle during mantled gneiss dome formation is well illustrated in the case of the Chester Dome. The unconformity is conspicuous on the western side of the Green Mountains (biotite-grade Paleozoic metamorphism) but is partly obliterated on the eastern side (garnet-grade). Brace (1958) has described the mechanism by which structures in the older rocks were rotated towards the plane of the unconformity during later deformation. Most of this adjustment occurred in a 3000 foot zone immediately beneath the unconformity. This obliteration of the unconformity is essentially complete in the Chester Dome (kyanite-staurolite grade) (Thompson, 1950). Here too, the adjustment was most severe at the flanks of the dome, where stratification and foliation in the older rocks precisely parallel those in the mantling rocks. At one locality mapped by the author about 1500 feet below the unconformity (Appendix 2, Locality 1), thin amphibolite layers show tight reversed-sense drag folds probably related to differential upward motion of the less dense core gneisses relative to the mantling strata. Nearer the contact these layers are drawn strictly parallel to the contact. At another locality (Appendix 2, Locality 2) about 3000 feet below the western contact,shallow-dipping amphibolite layers probably reflect the attitude of the Precambrian compositional layering. These have been partly dissected into the plane of a younger foliation parallel to the dome margins, and in other parts of the same roadcut compositional layering has been completely translated into the younger plane.

Some of the most interesting age measurements were made at Buttermilk Falls about one mile west of the unconformity on the east side of the Green Mountains. At this locality veins of quartzo-feldspathic pegmatite cut interlayered schist and quartzite of the Mt. Holly Complex. Rb-Sr ages on muscovite from the pegmatite vary with the size of the crystal selected for analysis and with position even in a single crystal (Figure 13). Sample GBF-1-mus is the center plate of a large, slightly deformed muscovite book which was cut into concentric rings labeled from I on the outside to IV in the center; sample GBF-1-mss is taken from the center of a smaller book; and sample GBF-1-m is an aggregate of fine-grained muscovite separated from a crushed sample of the pegmatite. All of the samples were removed from a single 20kg sample of the pegmatite. The data in Figure 5 show that the concentration of Rb is essentially constant throughout the large book, but that the concentration of normal Sr increases sharply in the outer rim. The concentration of radiogenic Sr ($Sr\ 87^{*}$) decreases outward from the center as does the measured age. An analogous effect is apparent in the smaller samples. There appears to have been major movement of common Sr from the surroundings into the outer parts of the crystals, and the data further indicate that the change in the measured ages is due to loss of radiogenic Sr rather than gain of Rb. A similar effect has been reported in biotite from a contact-metamorphic zone by Hart (1964). The data suggest that the pegmatite was emplaced at least 1050 million years ago. This age may be compared with Rb-Sr ages of 1035 ± 20 m.y. measured on pyroxene-hornblende granites in the

BUTTERMILK FALLS, VERMONT
MUSCOVITE SAMPLES FROM PEGMATITE



Sample	AGE 10^6 years	Sr^{87*}	Sr^{86}	Rb^{87}	$\frac{Rb^{87}}{Sr^{86}}$	$\frac{Sr^{87}}{Sr^{86}}$	$\frac{Sr^{87*}}{Sr^{87}}$
		(10^{-8} moles / gram)					
GBF-1-mus IV	1072	4.71	0.283	314	1108	17.365	95.9
GBF-1-mus III	1037	4.58	0.231	316	1367	20.576	96.5
GBF-1-mus I	1021	4.16	1.82	291	159.9	2.994	76.3
GBF-1-mss	998	4.45	0.418	319	762.9	11.366	93.8
GBF-1-m	928	2.74	1.73	216	122.0	2.292	69.1

Sr^{87*} = radiogenic Sr^{87} $\left(\frac{Sr^{87}}{Sr^{86}}\right)_{COMMON} = 0.708$

$\lambda\ Rb^{87} = 1.39 \times 10^{-11}$ year^{-1} $\left(\frac{Sr^{86}}{Sr^{88}}\right)_{NORMAL} = 0.1194$

Figure 13. Rb-Sr data for muscovite samples from pegmatite at Buttermilk Falls, Vt.

eastern part of the Adirondack Massif (Hills and Gast, 1964, p. 759).

Whole-rock Rb-Sr ages on the schists at Buttermilk Falls are puzzling. A 720 m.y. isochron with an initial Sr^{87}/Sr^{86} intercept of 0.708 (assumed value) can be passed through the data for samples GBF-7, -8,-9, and -10 (Table 2, Figure 14), yet the schists are intruded by billion year old pegamatite. There is nothing in the systematics of the data on the pegmatite muscovites to suggest that their measured age exceeds the age of emplacement of the pegmatite, so the whole-rock ages of the schist samples must have been systematically lowered. This effect is probably related to the fact that the Buttermilk Falls locality is a small Rb rich enclave in a terrain characterized by generally low Rb/Sr ratios. It is possible that during the metamorphism the schists lost radiogenic Sr to the surroundings in direct proportion to the amount of muscovite they contain. Since the Rb/Sr ratio of the schist samples probably varies proportionally with the concentration of muscovite, such a mechanism might account for the linearity of the array of data. The data demonstrate that a somewhat linear array of data on a strontium evolution diagram is not sufficient evidence for the validity of a measured age.

A small body of granite occurs on Dry Hill about seven miles north of Ludlow, Vermont. The granite intrudes gneiss of the Mt. Holly Complex but is truncated at the unconformity (Doll and others, 1961). It lies wholly within the zone of severe structural adjustment beneath the unconformity. Whole-rock samples have suitable Rb/Sr ratios for dating, but have been badly disturbed. Three whole-rock samples of the granite (Table 2; samples GDH-2,-3,and -4)

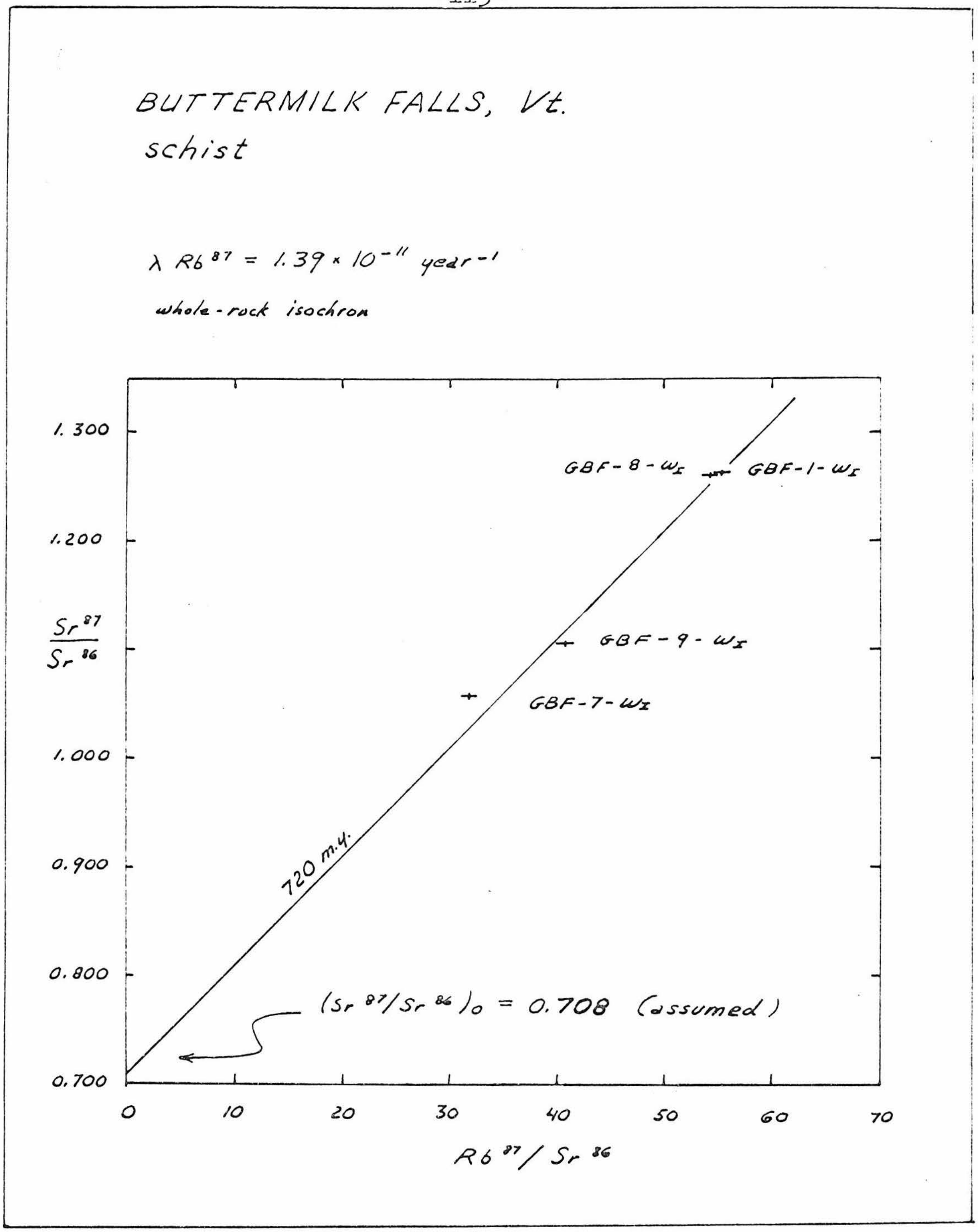


Figure 14. Strontium evolution diagram for schist samples, Buttermilk Falls, Vt.

have apparent ages (calculated assuming initial Sr^{87}/Sr^{86} = 0.708) of 1054 m.y., 1355 m.y., and 855 m.y. respectively. The spread in apparent ages is probably caused by failure of the whole-rock samples (average size about 10 kg) to behave as closed systems during garnet-grade Paleozoic metamorphism. The age of emplacement of the granite is probably within the range of apparent ages encompassed by the reported samples.

The core gneisses of the Chester Dome are characterized by very low Rb/Sr ratios. The average Rb/Sr ratio for fifteen samples of core gneiss is 0.14 with a range from 0.064 to 0.703 as measured by X-ray fluorescent spectrometry. Three samples duplicated by isotope dilution agreed with the X-ray data to within 10 percent. The samples are representative of the exposed parts of the Chester Dome. It is interesting to note that after 1000 million years the Sr^{87}/Sr^{86} ratio of material of this average composition would increase by only 0.006. Three whole-rock samples from a single locality in the Chester Dome (Table 2, samples CSG-1,-4, and -6) with Rb^{87}/Sr^{86} ratios less than 0.1 have Sr^{87}/Sr^{86} ratios ranging from 0.702 to 0.706. Since there is this much uncertainty in the composition of initial Sr in the dome rocks it is impossible to calculate even approximate ages from the samples available. A Precambrian lead-alpha age (900 m.y.) was measured by Faul and others (1963, Sample Vt-12) on zircon separated from a sample of the Chester core gneiss.

Several small intrusive bodies cut the core gneiss and lower mantle of the Chester Dome. The rocks are composed of albite, quartz, biotite, and muscovite with only traces of potassium feldspar and are thus similar to trondhjemites. One of these, the Baltimore Stock (Appendix 2, Locality 3) contains rotated xenoliths of the Gassetts Schist and other units of the lower mantle of the dome (Thompson, 1950). The assemblages in these xenoliths indicate the trondhjemites formed after the major metamorphic episode, but foliation within the trondhjemites indicates that some metamorphism continued after their intrusion. The trondhjemites may have formed at the same time as the doming of the core gneiss. They are texturally distinct from the core gneiss and probably were not formed in place by remobilization. They may however have formed by partial fusion of the core gneiss at greater depth. Because they are easily distinguished from the older gneisses and clearly intrude them no geologist has seriously argued from their presence that the entire core of the Chester Dome is an intrusive body younger than the mantling strata.

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