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METAMORPHISM OF LOWER PALEOZOIC ROCKS IN THE VICINITY OF THE TACONIC RANGE IN WEST-CENTRAL VERMONT[1]

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ABSTRACT

Rocks from the Taconic Range and adjacent areas, near Castleton in west-central Vermont, have been studied for their mineral assemblages. Most of the rocks are argillaceous or arenaceous, of Early Cambrian to Middle Ordovician age, and are from the Taconic allochthone. A number of samples, including carbonates, are from the Lower Ordovician to Middle Ordovician rocks of the surrounding autochthonous marble belt. Regional metamorphism was later than the emplacement of the Taconic allochthone; the metamorphic grades in the Taconic sequence cut across structures, and conform to the trend in the surrounding autochthone. The grade decreases from east to west, and the rocks range correspondingly from phyllites to slates and shales.

Study of the mineralogy of the samples made use both of the *x*-ray diffractometer and the petrographic microscope. Two chlorite samples, one from a slate and one from a phyllite and a chloritoid sample from a phyllite, in addition to four rock samples, were chemically analyzed.

The most significant mineral assemblages found are:

(1) muscovite-chlorite-quartz
(2) muscovite-paragonite-chlorite-chloritoid-quartz
(3) muscovite-chlorite-chloritoid-epidote-hematite-magnetite-quartz
(4) muscovite-stilpnomelane-chlorite-albite-microcline-quartz
(5) muscovite-biotite-quartz-graphite
(6) calcite-dolomite-chlorite-muscovite
(7) calcite-zoisite-chlorite

The prevalence of assemblage (1), with its simple mineralogy, indicates that chlorite may be variable in the (Fe, Mg)/Al ratio in addition to the Fe/(Fe+Mg) ratio. This is borne out by the chemical analyses; the analyzed chlorites are among the highest reported in their Al/(Fe, Mg) ratios.

The basal spacings of coexistent muscovite and paragonite in assemblage (2), found in the phyllites, are 9.97(5) Å and 9.62(3) Å, respectively, indicating very limited solid solution between these phases at this metamorphic grade. No paragonite has been found in the slates; it is suggested that paragonite may possess a lower limit of stability, with the more hydrous assemblage, kaolinite plus albite, taking the place of paragonite plus quartz.

Rocks relatively rich in Al contain chloritoid (2,3), and the chloritoid-chlorite association is common even in slates. The chlorite is always more Mg-rich although the analyzed chloritoid also shows significant solid solution of Mg for Fe. The only sodic phase found with chloritoid is paragonite. Rocks relatively poor in Al do not have chloritoid but contain stilpnomelane (4), in addition to microcline which is incompatible with high-Al phases. The location of stilpnomelane on the low-Al part of the A-F-M diagram is based on regarding all Fe‴ in stilpnomelane as subsequently oxidized Fe″; this is justified by an analysis of the available chemical data on this mineral. Stilpnomelane- and chloritoid-bearing as-

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semblages are mutually incompatible in the area; that the control is chemical composition rather than physical conditions of metamorphism is borne out by field mapping of the lithologic units. The stilpnomelane-microcline-chlorite assemblage rules out the chlorite-Fe biotite pair, which in fact is not found in the area although biotite does occur (5).

Data pertaining to phases in the A-C-F diagram are scanty. The mineral assemblages show, however, that chlorite may be the first mineral to form upon metamorphism of a quartzose and aluminous dolostone (6). The presence of the zoisite-calcite pair (7) indicates that the calcite-kaolinite pair, found in sedimentary rocks, ceases to be stable in the rocks of this area.

The analysis of the mineral assemblages in the Castleton area indicates that these rocks by and large have attained chemical equilibrium during metamorphism. The assemblages obey the Phase Rule, and no incompatible mineral pair has been found. Textural data on the rocks also suggest extensive recrystallization has occurred. Both H_2O and CO_2 may be treated as mobile components without violating the Phase Rule. Evidence, however, is inconclusive on the status of O_2.

Introduction

In the course of geological mapping at the north end of the Taconic Range and adjacent areas around the town of Castleton in west-central Vermont (Zen, in preparation), a large number of specimens of metamorphosed sedimentary rocks of Cambro-Ordovician age were collected. These rocks are predominantly argillites, but also include quartzites, graywackes, arkoses, and carbonates. Owing to regional metamorphism, the argillites now range from phyllites or fine-grained schists in the Taconic Range along the east side of the area, where the grade of metamorphism is the highest, to virtually unmetamorphosed shales in the Champlain Valley to the west. The other lithologic types show corresponding mineralogical and textural changes. With the stratigraphic succession and the nature of the geological structure fairly well known, it has become possible to study the metamorphism of these rocks.

The only extensive petrographic study of rocks of the Castleton area is that of T. N. Dale (1898). Dale made detailed microscopic studies of the slates, and published a number of chemical analyses of rocks from this area, as well as one mineral analysis, that of a pyrophyllite. No correlation was made by Dale between the results of chemical and microscopic investigations, however; and so the former are not very useful for modern petrological discussions.

In this paper, the observed mineralogical and petrographic details of the various minerals or mineral groups are first given, supplemented by chemical data. The mutual relationships of these minerals are then considered in the second section. Instead of considering each individual rock type as such, however, the various mineral assemblages are considered in terms of ternary or pseudoternary phase diagrams, and the mutual relation of the various mineral assemblages discussed in the light of these diagrams. In this manner, it is hoped, the dependence of the observed assemblages on the bulk chemical composition, for a given metamorphic environment, will be emphasized. The phase-diagram approach also makes it more easy to apply the principles of physical chemistry to the study of the mineral assemblages, such as predicting the direction of changes upon given changes in the environment, and understanding the mutual stability relationships of the mineral phases in a given assemblage. Lastly, the phase diagram approach, in one form or another, has been traditional for the study of metamorphic rocks of higher grades (see, for instance, Turner, 1948; Thompson, 1957; Korzhinskii, 1959); it is hoped that the present study may be easily integrated with these others in its present form.

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Resume of Regional Geology

As the full report on the geology of the Castleton area will be published elsewhere (Zen, in preparation; 1959b), only a brief summary is given here. The following is a simplified description of the stratigraphic succession in the area, the units being arranged from younger to older.

Lower to Middle Ordovician: Red, green, gray, and black, hard, quartzose slates, with interbedded fine, cherty quartzites, massive ankeritic quartzites, and thin, ribbony limestones. Some of the red slates are soft and of commercial value. The name "Poultney River group" is used informally for this group of rocks.

Lower Cambrian:

1. West Castleton formation. Black slate, with black, fossiliferous limestone near its base. The slate ranges from soft, fissile and pyritiferous to hard and sandy.
2. Bull formation. Mainly the Mettawee slate facies, consisting largely of purple,

green, and gray slate and argillite. Included also in the Bull formation are a fossiliferous intraformational conglomerate and a vitreous orthoquartzite near the top of the formation; a graywacke, the Zion Hill quartzite, near the middle of the formation; and the Bomoseen graywacke, a massive, olive-drab, areally extensive facies in the middle and lower part of the formation.

3. Biddie Knob formation. A purple to green, highly aluminous, chloritoid- and paragonite-bearing slate or phyllite.

These rocks, together, constitute an allochthonous unit, originally deposited perhaps at the site of the present Green Mountain anticlinorium. They were moved into the present position as a series of imbricate thrust plates, each of which is internally intricately and recumbently folded. The emplacement of the allochthone was an Upper Trenton event. The regional metamorphism was later superimposed on the complex regional structure, and has simple regional patterns, conforming to the grade and trend in the surrounding autochthonous units. The metamorphic grade generally decreases from east to west, so that, while at the eastern edge of the area the argillaceous rocks are phyllites and the carbonates contain actinolite and zoisite, at the western margin of the map area the rocks are barely metamorphosed.

The place names referred to in the text are located, to the nearest 1′ of latitude and longitude, in Appendix I at the end of the paper.

METHOD OF STUDY

The slates of the Castleton area are commonly so fine-grained that microscopic examinations do not always yield definitive results. Because of this fact, considerable reliance has been placed on *x*-ray methods of mineral identification. Due to the sensitivity of the results of such studies to the laboratory procedure, a brief description of the method used follows.

A Norelco high angle Geiger-counter diffractometer, equipped with Cu tube and Ni filter, was used in most of the work. The sample was ground fine and molded into standard aluminum holders by the use of a hand press. Such a method gives preferred orientation to inequidimensional minerals, particularly the platy minerals; however, for the purpose of identification of phases this fact presents no problem. For routine work, the following instrumental settings are used on the diffractometer: goniometer speed, 1° 2θ per minute, chart speed, $\frac{1}{2}''$ per minute; divergent, scatter, and receiving slits, 1°, 1°, and 0.003″, resp.; scale factor, multiplier, and time constant, 8, 1, 4, or 4, 1, 4, resp. The resulting charts are compared with prepared standards and the minerals thus identified. For careful measurements of line positions, a goniometer speed of $\frac{1}{4}°$ 2θ per minute is used with quartz added as an internal standard.

A simple scheme of mineral identification by the *x*-rays such as outlined above does not, of course, always yield clear-cut results. Despite the existence of such reference data as the A.S.T.M. file, the identification of an unknown phase often rests, in the last analysis, on evidence not intrinsic to the *x*-ray records. For instance, a study of the literature on the mineralogy of similar rocks may be helpful; or the thin section might yield clues to the mineral identities. One of the major drawbacks to the *x*-ray method is the possibility of line interference by several different phases in the sample; however, in most cases the combination of lines, rather than a single line, is used for identification, and such combinations are generally enough to define a phase. Nonetheless, the microscope has to be depended upon almost exclusively for the detection of minute amounts of a mineral, and for the study of the textural relations of the minerals. In all, about 200 thin sections have been examined.

MINERALOGY

Chlorite

Petrography and association. Chlorite is a common mineral in the rocks of the Castleton area; it is nearly ubiquitous in the argillaceous rocks and has been found in association with every other mineral except biotite. It occurs either in the bulk of the rock, or as irregular vein fillings, by itself or in association with quartz, quartz and albite, or quartz, albite, and carbonate.

Chlorite appears microscopically either as tiny scales, about 10μ in long dimensions, or as randomly oriented microporphyroblasts up to 0.5 mm. in size. As microporphyroblasts, it is commonly intergrown with muscovite and occasionally with stilpnomelane (?). The existence of chlorite in a specimen is easily checked by the *x*-ray diffraction pattern. The (00*l*) reflections are generally sharp to the fifth order. The only possible source of confusion comes from the 7 Å clay minerals such as kaolinite, but the latter lacks the reflection at about 14 Å.

Chemical analysis. Data on the composition of chlorites from low-grade rocks are scarce; yet such data are essential in discussions of the phase relations of the host rocks. Two chlorites, accordingly, have been separated and analyzed, and a third one computed from a rock analysis of known mineralogy. The two analyzed chlorites are given in Table 1, together with their optical properties and calculated formulae, taking the number of oxygens per formula as 18. Chlorite no. 197-2 was from a fracture filling in a soft, green slate of the Bull formation on the east side of Crystal Ledge ½ mile north of Bull Hill, and was virtually free of impurities, the small amount of alkali in the analysis having been computed as white mica, a trace of which was visible under the microscope. Chlorite no. 212-2A was from a paragonite phyllite found in the vicinity of Ben Slide north of Herrick Mountain, and the analyzed sample had 2% of quartz and 1.5% of muscovite and paragonite according to a point-count of the analyzed material. These impurities have been corrected for, but the chlorite composition is necessarily less reliable than no. 197-2.

TABLE 1. MINERAL ANALYSES

(Booth, Garrett and Blair, analysts)

Sample number	197-2 (chlorite)	212-2A (chlorite)	213-1A (chloritoid)
SiO_2	23.62	25.92	25.18
TiO_2	0.04	0.13	0.40
Al_2O_3	23.04	23.74	39.48
Fe_2O_3	2.20	1.70	2.00
FeO	31.18	27.67	22.26
MgO	8.64	8.66	1.25
MnO	0.15	0.30	1.32
CaO	none	0.36	0.06
Na_2O	0.12	0.30	0.15
K_2O	0.03	0.27	0.43
H_2O (1150°C)	11.00	10.65	7.30
H_2O (105°C)	0.20	0.00	0.06
Total	100.22	99.70	99.89
Optical properties: (sodium light)			
α	1.640 ± 0.001	-----	1.720 ± 0.002
β	1.642 ± 0.001	1.636 ± 0.001	1.722 ± 0.002
γ	-----	same	1.725 ± 0.002
2V	(-?) small	0°	(+) moderate
	X ⊥ (001)	-----	Y Λ (001) ≈ 20°
Pleochroic formula	Y = Z = deep green X = light yellow	Y = Z = deep green X = straw yellow	X = Y = indigo-green Z = straw yellow
Anomalous extinction?	blue	blue	bronze to lavender
			polysynthetic twinning with composition plane (001)
X-ray data:	d (001) 14.113 ± 0.002A	d (001) 14.113 ± 0.002A	d (001) 17.801 ± 0.002A

Calculated formula:

Chlorite no. 197-2

(a) All Fe‴ grouped with Fe" in the calculations

$$\underbrace{(Fe_{3.08}Mg_{1.43}Mn_{0.01}Al_{1.46})}_{5.98}\underbrace{(Al_{1.46}Si_{2.53})}_{3.99}O_{18}H_{8.07}$$

$(Fe'' + Mn)/(Fe'' + Mn + Mg) \approx 0.68$

(b) All Fe‴ grouped with Al in the calculations

$$\underbrace{(\underbrace{Fe''_{2.88}Mg_{1.42}Mn_{0.01}}_{4.31})(\underbrace{Al_{1.36}Fe'''_{0.18}}_{1.54})}_{5.85}(\underbrace{Al_{1.55}Si_{2.52}}_{4.07})O_{18}H_{8.03}$$

$(Fe'' + Mn)/(Fe'' + Mn + Mg) \approx 0.67$

$(Fe_{total} + Mn)/(Fe_{total} + Mn + Mg) \approx 0.68$

(c) According to the method of Brindley and Gillery (1956):

$$(Fe_{4.0}Mg_{0.7}Al_{1.3})(Al_{1.3}Si_{2.7})O_{18}H_8$$

Name [Hey, 1954, p. 280]: ripidolite

Chlorite no. 212-2A (corrected for quartz and sericite)

(a) All Fe‴ calculated as Fe"

$$\underbrace{(Fe_{2.85}Mg_{1.51}Mn_{0.03}Ca_{0.05}Al_{1.47})}_{5.91}\underbrace{(Al_{1.47}Si_{2.52})}_{3.99}O_{18}H_{8.08}$$

$(Fe + Mn)/(Fe + Mg + Mn + Ca) \approx 0.65$

(b) All Fe‴ calculated as Al

$$\underbrace{(\underbrace{Fe''_{2.71}Mg_{1.51}Mn_{0.03}Ca_{0.05}}_{4.30})(\underbrace{Al_{1.38}Fe'''_{0.15}}_{1.53})}_{5.83}(\underbrace{Al_{1.53}Si_{2.53}}_{4.06})O_{18}H_{8.10}$$

$(Fe'' + Mn)/(Fe'' + Mg + Mn + Ca) \approx 0.62$

$(Fe_{total} + Mn)/(Fe_{total} + Mg + Mn + Ca) \approx 0.65$

Name [Hey, 1954, p. 280]: ripidolite

Chloritoid no. 213-1A (all alkali computed as mica, titania as rutile)

(a) All Fe‴ calculated as Fe"

$$(Fe_{0.88}Mg_{0.08}Mn_{0.05}Ca_{0.00})Al_{1.94}Si_{1.00}O_7H_{2.10}$$

$(Fe + Mg)/(Fe + Mn + Mg) \approx 0.92$

(b) All Fe‴ calculated as Al

$$(Fe''_{0.82}Mg_{0.08}Mn_{0.05}Ca_{0.00})(Al_{1.94}Fe'''_{0.07})Si_{1.00}O_7H_{2.10}$$

$(Fe'' + Mn)/(Fe'' + Mn + Mg) \approx 0.92$

$(Fe_{total} + Mn)/(Fe_{total} + Mn + Mg) \approx 0.92$

In Table 2 is presented the rock analysis for specimen 212-2 from which, knowing the chlorite composition, the compositions of the white micas can be estimated.

In Tables 1 and 2 are presented the analysis for rock no. 213-1, and

TABLE 2. ROCK ANALYSES

(Booth, Garrett, and Blair, analysts)

Sample number	212-2	213-1	392-1	648-1
SiO_2	42.42	39.41	49.46	78.86
TiO_2	1.10	0.66	1.02	0.70
Al_2O_3	33.89	31.21	25.36	8.85
Fe_2O_3	2.40	2.10	1.96	1.86
FeO	4.87	10.36	6.33	1.63
MgO	1.93	3.41	2.01	0.76
MnO	0.08	0.16	0.08	0.02
CaO	0.12	0.14	0.22	0.56
Na_2O	2.40	1.55	4.25	1.69
K_2O	5.07	4.01	4.14	2.37
H_2O (1000° C.)	5.80	6.54	4.90	1.97
Total	100.14	99.55	99.73	99.27
Modal mineralogy:	Muscovite Paragonite Chlorite Quartz Rutile	Muscovite Paragonite Chloritoid Chlorite Quartz Rutile	Albite Muscovite Chlorite Quartz Rutile	Muscovite Chlorite Stilpnomelane Plagioclase (albite) Microcline Quartz

Weight % norms (based on identified mineralogy):

(a) All Fe''' as Fe''	Muscovite	42.9	Muscovite	34.2	Muscovite	35.4
	Paragonite	29.6	Paragonite	19.2	Chlorite*	23.1
	Margarite	0.8	Margarite	1.0	Albite	36.6
	Chlorite	21.6	Chloritoid	8.7	Anorthite	1.1
	Quartz	4.0	Chlorite	31.2	Quartz	2.7
	Rutile	1.1	Quartz	5.1	Rutile	1.1
			Rutile	0.7		
(b) All Fe''' as Al	Muscovite	43.2				
	Paragonite	29.8				
	Margarite	0.8				
	Chlorite	17.4				
	Quartz	7.6				
	Rutile	1.1				
(c) *X*-ray mode					Chlorite	23
					Muscovite	30
					Plagioclase	43
					Quartz	3†
					Rutile	1†

Calculated chlorite composition: (Number 213-1)

(a) All Fe''' calculated as Fe'', both in chlorite and chloritoid
$(Fe+Mn)/(Fe+Mn+Mg+Ca) \approx 0.62$

(b) All Fe''' calculated as Al, both in chlorite and chloritoid
$(Fe''+Mn)/(Fe''+Mg+Mn+Ca) \approx 0.51$
$(Fe_{total}+Mn)/(Fe_{total}+Mg+Mn+Ca) \approx 0.56$

(c) According to Winchell and Winchell (1951, p. 385) and the "$\frac{1}{3}$ substitution" formula:
$(Fe+Mn)/(Fe+Mg+Mn) \approx 0.67$

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* Chlorite based on "$\frac{1}{3}$ substitution" formula.

† Presence assumed in the computations.

for the chloritoid separated therefrom. The rock is from a locality about 500 feet due east of the locality for specimen 212-2. From the composition of the chloritoid, the Fe/(Fe+Mg+Mn) ratio in the coexistent chlorite has been computed, as given in Table 2.

Calculation of the chlorite formula. The calculated composition of a chlorite depends on the structural nature to be assigned to the ferric iron given in the chemical analysis. On the one hand, it might be substituting for aluminum, and must be reckoned therewith; on the other hand, it might be ferrous iron that has undergone oxidation after the formation of the chlorite. The corresponding chlorite formulae constitute limits of variation between which the true composition must lie. These calculated formulae are presented in Table 1.

Discussion. Both chlorite analyses show considerable excess of Al over the ideal Pauling formula of 2 Al's per 18 oxygen (Pauling, 1930). This indicates that an added compositional variation exists in chlorite, in that the Al^{iv}/Si ratio is variable, a fact familiar to the mineralogists (Winchell, 1926; Orcel, 1927; Brindley and Robinson, *in* Brindley, 1951; Hey, 1954).

Thompson (1952, lecture notes) suggested that, in order to achieve local valence saturation in the chlorite structure, $\frac{1}{3}$ additional Al per 18 oxygen, in each of the octahedral and tetrahedral positions of the talc-type layers, may be expected. The calculated formulae just given exceed even this value. The following schemes of atomic substitution are tentatively suggested to account for the excess Al.

First there is the possibility of a random mixed-layer structure between talc-type and pyrophyllite-type layers sandwiched in between the brucite-type layers. Against this hypothesis, however, may be cited the lack of evidence for the completely analogous mixed-layer structure between biotite and muscovite (Yoder and Eugster, 1953, p. 67); this lack of solid solution may be explained by the strain that would be induced in the lattice due to differences in the *a* and *b* dimensions of the dioctahedral and trioctahedral layers.

Second, the excess Al in chlorite may be accommodated, again with local valence balance, by the simultaneous replacement of a divalent atom in the brucite-type layer and a silicon atom in the tetrahedral position to yield a chlorite having the following formula as a component:

$$(Fe, Mg)_2Al(OH)_6(Fe, Mg)_2AlAl_2Si_2O_{10}(OH)_2.$$

Indeed, Nelson and Roy (1958, p. 712) have found experimentally that a complete solid solution apparently exists between chlorite with the above formula and with the Pauling formula.

Finally, it is possible to accommodate some extra Al within the octahedral layer by the simple scheme

$$2Al \rightarrow 3(Fe, Mg).$$

This scheme is impossible to differentiate, by chemical data alone, from the first one; it should, however, be limited in extent since it implies strong local bond-strength variations leading to a less stable structure.

Problems in the identification of chlorite. The problem of determination of the chemical composition of chlorites by means of physical measurements has long interested the mineralogist. Hey (1954) recently summarized the data to date and attempted to establish regression equations between the composition and the indices of refraction, cell dimensions, and the bulk density of the chlorites. Many other attempts also exist at relating some specific compositional variation with certain cell-dimension changes (see, for instance, Brindley and Robinson, *in* Brindley, 1951; Mikheev, 1953; Nelson and Roy, 1958), and Brindley and Gillery (1956) recently advocated the use of *x*-ray line intensities for the same purpose.

Hey's method is based on the assumption of the existence of five independent compositional variables, which he chose to be Si, Fe''', Fe^{total}, Mn, and Cr. To use his relations, one must assume (1) that there is no hydrogen-deficiency in the structure, and (2) that ferric iron occupies either only 4-fold or only 6-fold positions, for otherwise there is no unique solution to the Mg and Al contents in the mineral.

The defects of earlier attempts at the determination of chlorite composition by *x*-ray have been summarized by Brindley and Robinson (*in* Brindley, 1951). The method of Brindley and Gillery, based on a combination of cell-dimension and intensity studies, so far seems to give only qualitative information (1956, p. 182, fig. 3), as is shown by their own results. Chlorite no. 197-2 was studied this way and the results are incorporated in Table 1 for comparison with direct chemical analysis.

In the following discussions, unless otherwise specified, a "$\frac{1}{3}$ substitution" formula is assumed for chlorite. The $FeO/(FeO+MgO)$ ratio is then determined by using the chart given by Winchell and Winchell (1951).

Chloritoid

Petrography and association. Chloritoid has been reported from low-grade rocks of various areas; as examples may be cited the Tintagel district (Tilley, 1925), Cornwall (Phillips, 1928), the Gotthard area (Niggli, 1912), southeastern Pennsylvania (Agron, 1950), and the Berkshire Hills of Massachusetts (Balk, 1953). In the Castleton area, chloritoid occurs abundantly not only in the phyllites of the Taconic Range, but even in the purple and green slates in the vicinity of Lake Bomoseen.

On the *x*-ray diffractometer, chloritoid is distinguished by a sharp line pair, corresponding to the (004) and (006) reflections, at about 20° and 30° 2θ, respectively. Microscopically, chloritoid typically occurs as lath-shaped, euhedral, polysynthetically twinned crystals, about 0.1 mm. in length. The crystals are randomly oriented in the rocks, and commonly cluster together in sheaves. The high relief, low birefringence, and segmented appearance, coupled with pleochroism from indigo to yellow, are diagnostic. Chloritoid commonly intergrows with muscovite, occupying the centers of the "sandwiches." The effect is most striking in hematitic rocks, as the compound porphyroblasts show up as clear areas in a pasty, translucent background. The size of the porphyroblasts, about 0.1 mm., affords a minimum measure of the distance of material transport during recrystallization.

In the Castleton area, chloritoid has been found to coexist with the following minerals: muscovite, paragonite, chlorite, epidote, hematite, magnetite, rutile, and quartz. Albite has not been found with chloritoid; this may be due in part to low sodium content of the aluminous pelites in the area so that no distinct sodic phase is formed; however, it may also reflect the low metamorphic grade, so that this mineral pair is unstable, a point to be considered later. Chloritoid has not been found with biotite in the present study, but the evidence is inconclusive due to the rarity of biotite. Hietanen (1941, p. 144, 146) reported chloritoid with albite and biotite in the Wissahickon schist in Pennsylvania.

Chemical analysis. Table 1 gives a chemical analysis of chloritoid from specimen 213-1, from a locality east of Ben Slide. The analyzed sample had 0.4% of sericite and 0.1% of quartz as impurities (point-count analysis) and was full of tiny rutile inclusions. In the calculations of chloritoid formula, therefore, all alkali has been taken as white mica and all titanium oxide as rutile. The resulting formulae are given in Table 1, both when the analyzed Fe''' is taken as original Fe''' and when it is taken as oxidized Fe''. Both formulae are so close to the ideal formula of $(Fe, Mg)Al_2SiO_7H_2$ that they are equally likely. The analysis shows that 8% of the octahedral, divalent position in chloritoid is taken by magnesium. This is a minimum measure of the amount of Mg that chloritoid can dissolve in an environment represented by the rocks of this vicinity, and shows that even at such low metamorphic grades chloritoid cannot be taken as a purely ferrous mineral.

Muscovite and Paragonite

Muscovite and paragonite occur in slates and phyllites typically as microscopic flakes, about 20μ long (in the *ab* plane). They are distinguished from other platy minerals by the relatively high birefringence and lack of color and pleochroism. The flakes are commonly oriented parallel to one of the rock cleavages.

Muscovite also occurs as bulbous microporphyroblasts about 0.1 mm. across, cross-cutting the rock cleavage directions and commonly interleaved with chlorite and less commonly with chloritoid. This textural feature demonstrates that the minerals are the results of recrystallization, and not detrital in origin.

On the *x*-ray diffractometer, the pattern of muscovite is distinctive. The most useful reflections are the strong basal lines; the exact *d* spacings of which are a measure of the amount of sodium (and possibly calcium) in solution in muscovite. The one possible source of confusion is biotite, which has the same basal spacing. Iron-bearing biotite is generally pleochroic and readily identified with the microscope, though pure Mg biotite (phlogopite) could escape notice. For most of the rocks, however, there is commonly an iron-bearing chlorite, hence it is not likely that any coexisting biotite could be purely magnesian.

Paragonite is microscopically indistinguishable from muscovite in the rocks studied. Rosenfeld (1956, p. 144) reported differences in the 2V values and in refractive indices; the method is not adaptable to very fine-grained samples. On the *x*-ray diffractometer, however, the two phases are readily differentiated by their basal spacings. Although the strong 3.2 Å line of paragonite coincides with the (002) reflection of albite, the over-all pattern of paragonite is sufficient for its certain identification.

Muscovite is found in the Castleton area in association with all minerals. In fact, rarely is there a rock that does not carry muscovite. In the majority of the rocks studied, this mineral represents the principal potassic phase; only a few rock types, notably the Bomoseen greywacke, a few slates of the Poultney River group, and a few limestones, carry a potassic feldspar as well as muscovite.

In contrast, paragonite is rare in rocks of this region and has not been found in the true slates. It is found in significant amounts only from a low outcrop at 2000′ level, about $\frac{1}{4}$ mile east of Ben Slide (specimen 212-2). The rock is a highly micaceous phyllite, containing also muscovite, chlorite, and a little quartz and rutile. The basal spacings of the two micas in this rock are as follows:

Paragonite, *d*(001)	9.62(3) Å
Muscovite, *d*(001)	9.97(5) Å

Traces of paragonite have also been found, in this area and near Biddie Knob, in the Biddie Knob formation, in coexistence with chloritoid as well as the above-mentioned minerals.

In recent years paragonite has become increasingly recognized, thanks to the *x*-ray diffractometer, as an important mineral in low- to medium-grade metamorphic rocks high in aluminum. The various mineral associations of paragonite have been described (Rosenfeld, 1956; Harder, 1956; Dietrich, 1956; Eugster and Yoder, 1955; Albee, 1957b). A study of the paragonite-muscovite pair has been published (Rosenfeld, Thompson, and Zen, 1958) and the mineral assemblages will be considered in a separate paper by the same authors. From the *x*-ray lattice spacing data, the paragonite and its coexistent muscovite from the Castleton area represent nearly pure sodic and potassic micas, respectively; and with the possible exception of the brammallite reported by Bannister (1943), the occurrence of paragonite at Ben Slide may represent the lowest-grade assemblage yet recorded.

Stilpnomelane and Biotite

Petrography and association. The micaceous mineral stilpnomelane has been found commonly in the Bomoseen greywacke member of the Bull formation in the Castleton area. It is found in the most easterly areas of the greywacke outcrop, just west of the main Taconic Range; it is also found in the most westerly area, at Mill Pond near Benson. Since the metamorphic grade generally decreases westward, this mineral appears to occur over the entire range of metamorphic grade in the Castleton area.

The Bomoseen greywacke is a fine-grained, olive-drab, poorly-cleaved rock with occasional grains of quartz, feldspar, muscovite, and rock fragments visible in the hand specimen. Stilpnomelane cannot be seen even with the aid of a hand lens.

Under the microscope, stilpnomelane is seen as tiny scales, generally less than 0.02 mm. in their long dimensions, interstitial to and wrapping around quartz and feldspar grains, as well as in separate sheaves. The scales may be interleaved with muscovite or chlorite, demonstrating that they are the results of recrystallization and not detrital in origin. The scales are strongly pleochroic, and resemble brown biotite closely except for the occasional segmented appearance and lack of mottling between crossed nicols, in agreement with Hutton's observations (1938, p. 182). Optical data on stilpnomelane and coexisting chlorite from one of the specimens are given on page 163. The basal spacing, 12.1 Å, agrees with those of Gruner (1944, p. 298) and Hutton (1938) and proves its identity. The 12.1 Å line is not shifted by treatment with ethylene glycol. On the basis of its pleochroic color and index of refraction, the stilpnomelane may be rather rich in ferric iron according to the data of Hutton (1938, p. 177, 182, 184, 187). This is in agreement with the rather high ferric-ferrous ratio (about 1:1 atomic ratio) in the rock analysis for specimen 648-1 (Table 2), which contains stilpnomelane.

The correlation of the indices of refraction of stilpnomelane with composition is uncertain. Hutton (1938, p. 187) and Winchell and Winchell (1951, p. 390) give regression charts. Winchell and Winchell, however, take into account only the effect of oxidation of iron and manganese, whereas Hutton assumes that ferric iron and aluminum have the same effect on the indices of refraction, which is probably not true.

Stilpnomelane has been found in association with the following minerals in the same rock: plagioclase (nearly pure albite by extinction-angle and $\Delta 2\theta$ measurements; the latter based on data of Tuttle and Bowen, 1958, p. 12), microcline, muscovite, quartz, chlorite, and hematite. Microcline and hematite are rare, but the associations with the other phases are common. These phase relations will be discussed in a later section.

Besides stilpnomelane, a number of rocks in the Castleton area carry a pleochroic, micaceous mineral, probably biotite or stilpnomelane, which because of its extreme rarity in the specimens cannot be definitely identified. Two characteristic appearances of the mineral(s) are described below.

1. Small (about 0.05 mm.) scales, in green, grey, and purple slates and greywackes, found as far west as the neighborhood of Glen Lake. These are pleochroic, with X = light green, Y = Z = dark green. The grains look like stilpnomelane. Due to rarity of grains, the indices of refraction have not been ascertained. For the same reason this mineral cannot be observed on the x-ray diffraction chart. The mineral association is quartz, albite, chlorite, muscovite, and hematite.
2. Tiny books, characteristically in graphitic black slates or quartzose nodules therein. Y = Z = light olive brown, X = light yellow. Between crossed nicols this mineral appears mottled. It is most probably biotite. Associated minerals are quartz, muscovite, and graphite.

Problem of stilpnomelane. Special interest is attached to the mineral association of stilpnomelane because of the rather uncertain composition and structure of this phase. The mineralogy has been summarized elsewhere (Hutton, 1938; Gruner, 1944); it seems certain that structurally this mineral is related to the micas. The chemical analyses of stilpnomelane (Hutton, 1938, 1956; Gruner, 1944; Kojima, 1944; Shirozu, 1955) are difficult to interpret because apparently the composition is independently variable along several coordinates, a point which, however, cannot be established. The chemical analyses, compared with the analyses of biotite and minnesotaite, consistently run relatively high in silica and calcium. The role of ferric iron in the analyses presents additional difficulties. To see in what manner these factors affect the field of composition of stilpnomelane, and in what ways the compositions may be related to those of iron biotite and minnesotaite, the best published chemical analyses are plotted in Fig. 1 under three different assumptions: (1) all of the ferric iron is original and structurally substitutes for Al; (2) all ferric iron represents oxidized ferrous iron of the original mineral;

and (3) ferric iron plays a distinct and independent role in the structure and therefore does not plot on the diagram. One could also make different plots by assigning to Ca different structural roles, for instance substituting for the alkalis, substituting for the divalent metal atoms, and having its own structural position and therefore not plotting on the diagrams at all. The amount of Ca is so small, however, that these different plots appear sensibly the same. In Fig. 1, Ca is ignored.

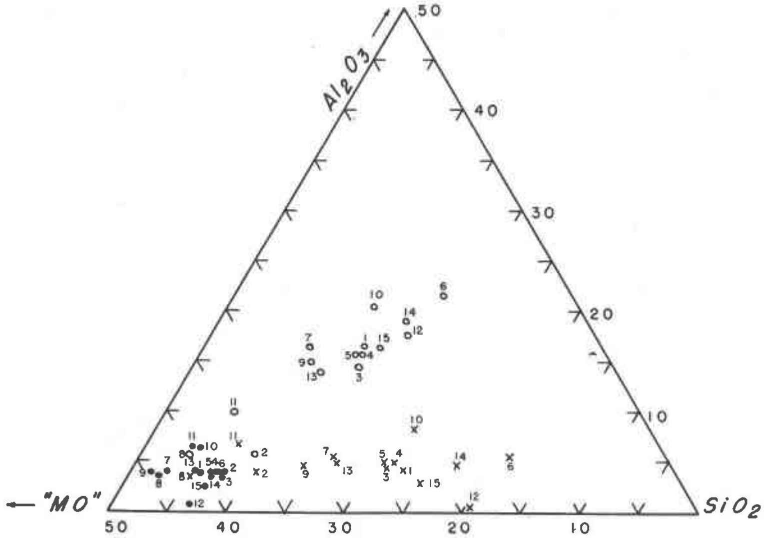

FIG. 1. SiO_2-Al_2O_3-"MO" plot of analyzed stilpnomelane.

Open circles: data analyzed according to assumption (1), using the following coordinates:

$Al_2O_3 + Fe_2O_3 - Na_2O - K_2O$; $FeO + MnO + MgO$; $SiO_2 + 2K_2O + 2Na_2O$

Solid circles: data analyzed according to assumption (2), using the following coordinates:

$Al_2O_3 - Na_2O - K_2O$; $FeO + MnO + MgO + 2Fe_2O_3$; $SiO_2 + 2K_2O + 2Na_2O$

Crosses: data analyzed according to assumption (3), using the following coordinates:

$Al_2O_3 - Na_2O - K_2O$; $FeO + MnO + MgO$; $SiO_2 + 2Na_2O + 2K_2O$

Sources of data:

Numbers 1 through 6, Hutton, 1938; p. 184, nos. B through G, resp.;
Numbers 7, 8, and 9, *ibid.*, p. 185, nos. B-2, C-1, and E-1, resp.;
Numbers 10 and 11, Kojima, 1944, p. 328, nos. I and II, resp.;
Number 12, Shirozu, 1955;
Numbers 13, 14, and 15, Hutton, 1956, p. 611, nos. A, B, and C, resp.

The results are interesting in that the points show enormous scatter except when ferric iron is treated as oxidized ferrous iron. Shirozu's analysis is the only one which may require calculation of Fe′′′ as Al, depending on the structural position assigned to Ca. It is interesting to note that, on the basis of microscopic evidence, Hutton (1938, p. 181) also suggested that ferric stilpnomelane may be due to later oxidation of the ferrous component.

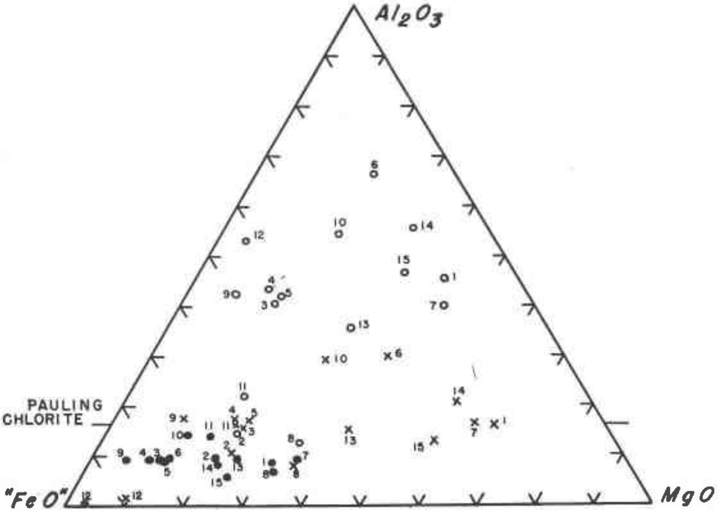

FIG. 2. A-F-M plot of analyzed stilpnomelane. The position of chlorite according to Pauling's ideal formula is indicated; biotite according to the ideal formula would plot along the base of the diagram.

Open circles: data analyzed according to assumption (1), using the following coordinates:

$$Al_2O_3 + Fe_2O_3 - Na_2O - K_2O; \quad FeO + MnO; \quad MgO$$

Solid circles: data analyzed according to assumption (2), using the following coordinates:

$$Al_2O_3 - Na_2O - K_2O; \quad FeO + MnO + 2Fe_2O_3; \quad MgO$$

Crosses: data analyzed according to assumption (3), using the following coordinates:

$$Al_2O_3 - Na_2O - K_2O; \quad FeO + MnO; \quad MgO$$

The numbers have the same significance as in figure 1.

Ignoring calcium, the results are also plotted on an A-F-M diagram, with (Al_2O_3-Na_2O-K_2O), FeO+MnO, and MgO as the apices, under the same set of three assumptions. The results are given in Fig. 2. Assumption (2) gives good coherence to the data. All the points plotted under assumption (2), significantly, fall below the $Al_2O_3/(Al_2O_3+FeO+MgO+MnO)$ line for chlorite according to the Pauling formula, and in fact fall between this and the line for biotite, due allowance being made for the "$\frac{1}{3}$ substitution" analogous to that for chlorite (Thompson, 1952, lecture notes). The stilpnomelane analyses, under assumption (2), are also plotted on the A-K-F(M) diagram, Fig. 5.

Carbonate

Two carbonates are found in rocks from the Castleton area: calcite and dolomite. Both occur as beds, lenses, and nodules, as well as subsidiary constituents in slates, quartzites, greywackes, and so forth. East of the Taconic Range in Whipple Hollow, the Ordovician limestones (Zen, 1959b) are sugary-textured marbles although the dolomitic beds may remain fine-grained; the average grain size decreases westward across the slate belt, paralleling the lowering of regional metamorphic grade in that direction, so that at the western margin of the map area the limestone rarely shows visible calcite grains in a hand specimen.

Calcite, under the microscope, generally is anhedral and tends to be considerably coarser grained than the coexisting dolomite, which commonly is in rhombs. The identities of these two carbonates have been checked by *x*-rays, making use of the strong and distinctive (10.4) reflections.

In the Castleton area, both dolomite and calcite have been found with the following minerals: quartz, muscovite, chlorite, albite, a potassic feldspar which is at least in some cases microcline, hematite, and possibly chloritoid. The association of microcline with carbonates occurs commonly in limestones, where the amount of argillaceous matter is low and thus microcline is presumably stable. The single occurrence of zoisite, in the marble at West Rutland, is associated with calcite.

Aside from quartz and the feldspars, chlorite and muscovite are the most common accessory phases in the carbonate rocks. Muscovite is commonly in euhedral, thin books, slender and lath-shaped in thin section and cross-cutting the carbonate grain-boundaries; chlorite may assume the same textural relations and is strongly colored due to its iron content. Less commonly, it is nearly colorless, with positive optical sign, and occurs as clots of delicate scales between carbonate grains, apparently as a magnesium-rich phase (Albee, 1957a, p. 80) resulting from reaction between the carbonate and an aluminum-rich phase during metamorphism. The textural relations also suggest that muscovite and chlorite are not metastable detrital grains.

Feldspar

Only two feldspars are known from the Castleton area: albite and potassic feldspar, some samples of which are definitely identified as microcline.

Albite is a fairly common accessory in slates and phyllites that do not carry chloritoid. It is the major mineral in many albite-phyllites, found sporadically along the contact of the green and black phyllites east of Taconic Range. As what appears to be detrital grains, it is a common mineral in greywackes and arkoses and in various limestones.

Albite in slates and phyllites is difficult to identify microscopically, as it is generally fine grained and untwinned. The albite from porphyroblastic phyllites is commonly also untwinned, typically occurring as round grains up to 0.5 mm. across, and showing, under the microscope, numerous helicitic inclusions of the mica and carbonaceous material of the matrix, indicating bodily rotation of the grains during growth. Detrital albite is generally polysynthetically twinned according to the albite law. Such albite, especially in greywackes, is commonly sericitized. The nature of the sericite is uncertain due to the very small amounts.

The composition of the albite is deduced from a combination of refractive index measurement, extinction-angle measurement according to the method of Michel-Lévy (when feasible), and *x*-ray lattice spacing measurement. The first two methods, especially the index of refraction, are sensitive to calcium content, which is found to be negligible. The potassium content is then estimated by the angular separation of the $(\bar{2}01)$ and the quartz (10.0) lines, using the regression relation now published by Tuttle and Bowen (1958, p. 13). These data are strictly applicable only to high temperature, synthetic feldspars, and show decreasing angular separation with increasing potassium content. Albite from the Castleton area shows this angular separation to a degree exceeding that of pure high-temperature albite ($\Delta 2\theta = 1.14°$), but agrees exactly with that of Amelia albite ($\Delta 2\theta = 1.18°$ according to the data of Smith, 1956, p. 52) which is generally accepted to have the low-temperature feldspar structure. The albite samples from the Castleton area, accordingly, are nearly pure $NaAlSi_3O_8$; this is true even of the albite in carbonate rocks. Combined with the observation of sericitization, it shows that even large detrital grains closely approached chemical equilibrium during regional metamorphism. That albite is the product of low-grade metamorphism of plagioclase feldspar is also demonstrated by the work of Coombs (1954).

In the Castleton area, albite is found to coexist with most other minerals. The exceptions are chloritoid, paragonite, and epidote (or zoisite). Imperfect sampling may account for the last two relations; the absence of the albite-chloritoid pair may be due to the low metamorphic grade.

A potassic feldspar is abundant in limestones in the Castleton area. In most cases it is definitely microcline, showing quadrille twinning; it is angular and commonly fragmentary, indicating a detrital origin. Microcline probably is present also as porphyroblasts in a few albite phyllites from the eastern slopes of the Taconic Range. Such porphyroblastic "microcline" shows at most simple Carlsbad twin, and is identified chiefly on the basis of its optical properties: 2V (−) large, $\gamma <$ balsam. *X*-ray evidence is inconclusive, as the amount of this is small, and the diffraction peak at 27.5° 2θ would be weak and interfered with by the rutile peak. Staining of the feldspars by sodium cobaltinitrite solution, according to the simplified technique of Chayes (1952), is equally inconclusive largely due to the fine grain size.

Microcline, showing quadrille twinning, occurs in the Bomoseen greywacke and occasionally, as detrital grains, in other greywacke and arkose beds. A potassic feldspar was also identified by *x*-rays in some of the Poultney River slates; but here the feldspar grains are too fine for microscopic identification.

Other Minerals

Pyrophyllite. Dale (1898, p. 191; 1900, p. 19) reported pyrophyllite from the phyllite of Ben Slide, north of Herrick Mountain. Enough sample was purified for a chemical analysis. Dale (1898, p. 191) described the mineral as "bright green . . . in thin scales." The rocks found today at Ben Slide are purple and green phyllite carrying muscovite, chlorite, chloritoid, quartz, and exceptionally paragonite. The writer has collected some 50 likely-looking samples from Ben Slide and studied them by *x*-rays. Not a trace of pyrophyllite has been found, and so its mineral association here remains unknown.

Epidote and zoisite. In the Castleton area, a mineral of the epidote group is found in a purple phyllite, specimen 29-4-1, from the west slope of Clark Hill, northeast of Herrick Mountain. It occurs as rare, large (about 0.4 mm. across), randomly oriented euhedral grains. Due to its rarity, however, the precise nature of the mineral cannot be determined. Associated phases are chloritoid, chlorite, muscovite, quartz, rutile, hematite, and traces of probable magnetite.

Zoisite is definitely found in a Beekmantown marble at West Rutland. It is colorless and euhedral, in slender prisms up to 3 inches long. Its identity is established by both *x*-ray and microscopic methods. The associated minerals are calcite and chlorite.

Quartz. Quartz is practically ubiquitous in the Castleton area. With few exceptions, it shows clear-cut extinction between crossed nicols. Especially in phyllitic rocks, it is commonly sufficiently recrystallized to give polygonal outlines in a thin section, with straight contacts between neighboring grains. The longest dimensions parallel the rock foliation. This relation shows that the grains have approached textural equilibrium through recrystallization.

Tourmaline. Tourmaline is a very common accessory in slates of the Castleton area. It characteristically forms euhedral crystals cross-cutting the rock cleavage. In some specimens the amount of tourmaline may be up to 1% of the rock by volume.

Sphene. Sphene is found only in specimen 29-4-1, which also carries epidote. Apart from rutile, it is the only titanium mineral found in the area.

Rutile. Rutile is the common titanium mineral found in the area. It characteristically forms minute needles randomly distributed in a thin section, and is microscopically discernible only under extreme magnification. Rutile is found in all assemblages in this area.

Iron oxides. Two iron oxides have been observed: hematite, found in the slates and much of the phyllites, and magnetite, rarely found in the slates (specimen 233-1-2), and more commonly in the phyllites of the Taconic Range. Hematite and magnetite are found together only in specimen 29-4-1, already described.

Hematite has been found to coexist with every known mineral in the Castleton area except paragonite. Under the microscope, it is readily recognized by the red color, high relief, and translucency; moreover, under high magnification the hexagonal outline of the basal plates can usually be seen, showing that it is not goethite.

Magnetite is recognized mainly in reflected light under the microscope, as very fine and rare grains in thin sections. Its identity therefore is not beyond doubt; the amount is commonly too small for checking with a magnet. It is observed in coexistence with the same minerals as hematite.

PETROLOGY

Mineral Assemblages

Before embarking on a discussion of the mutual relations of the minerals, it is convenient to list all the observed mineral assemblages. In the list that follows, no regard is paid to the relative *amounts* of the phases present, although differences in amounts may result in rocks of very different appearances and may even imply significantly different origins. For these gross lithological differences, the reader is referred to the field report (Zen, in preparation). The accessory minerals that might be described as the "heavy mineral suite" are also ignored. These minerals have not been studied as such and in all probability do not participate in any important way in the phase equilibrium relations to be considered in this part of the paper. Phases that are listed in parentheses are subsidiary in amount and may or may not be present in any given specimen.

1. Muscovite-chlorite-quartz (rutile, carbonate)
2. Muscovite-chlorite-paragonite-quartz (rutile)

3. Muscovite-chlorite-albite-quartz (rutile, carbonate)
4. Muscovite-chlorite-hematite-quartz (rutile, albite, carbonate)
5. Muscovite-chlorite-magnetite-quartz (rutile, albite)
6. Muscovite-biotite-quartz-graphite
7. Muscovite-stilpnomelane-chlorite-albite-quartz (rutile)
8. Muscovite-chlorite-stilpnomelane-albite-microcline-quartz-hematite (rutile)
9. Muscovite-chlorite-potassic feldspar-quartz
10. Muscovite-chlorite-chloritoid-quartz (rutile)
11. Muscovite-chlorite-chloritoid-paragonite-quartz (rutile)
12. Muscovite-chlorite-chloritoid-hematite-quartz (rutile)
13. Muscovite-chlorite-chloritoid-epidote-hematite-magnetite-quartz (rutile)
14. Muscovite-chlorite-albite-potassic feldspar-quartz (rutile, graphite)
15. Quartz
16. Calcite (quartz, albite, microcline, muscovite, chlorite, hematite)
17. Dolomite (quartz, albite, microcline, chlorite, muscovite)
18. Calcite-dolomite (quartz, albite, microcline, muscovite, chlorite)

The following assemblages are found in boudinage fillings:

19. Calcite
20. Quartz
21. Chlorite
22. Quartz-chlorite
23. Quartz-chlorite-albite
24. Quartz-chlorite-albite-ankerite

ROCK ANALYSES

In Table 2 are given four analyses of rocks from the Castleton area. The following brief descriptions and discussions may be helpful.

Rock no. 212-2. This rock, from which the chlorite 212-2A of Table 1 was extracted, is a highly micaceous, lustrous, medium-coarse grained, green-grey phyllite of the Biddie Knob formation. The outcrop from which the specimen was collected is at 2000′ elevation, about 1000 feet east of Ben Slide, and S75°W from the 1930′ knob west of Clark Hill. The rock contains, on the basis of *x*-ray data, muscovite, paragonite, chlorite, quartz, and rutile. Knowing the composition of the chlorite, it is possible to calculate both the composition and amounts of the white micas. It turns out, unfortunately, that the results of this calculation depend on whether the Fe‴ in the analysis is assumed to be oxidized Fe″ or if it is truly trivalent iron substituting for Al in the minerals. Accepting these two assumptions as limits, the calculated percentages of these minerals are given in Table 2. The β index of the muscovite, using Na light, was 1.604±0.001.

Rock no. 213-1. This rock, from which the chloritoid 213-1A of Table 1 was separated, is a green, friable, lustrous phyllite of the Biddie Knob formation, collected from a small northeast-trending ravine, at 1850′ level about 1500 feet east of Ben Slide. Besides chloritoid, the rock contains chlorite, muscovite, quartz, rutile, and some paragonite. Knowing the chloritoid composition, and assuming an ideal formula of $(Na,K)Al_3Si_3O_{10}(OH)_2$ for the white micas and the "$\frac{1}{3}$ substitution" formula for chlorite, the latter has been calculated to have the following alternative formulae, depending on the structural role of ferric iron:

(1) All Fe‴ calculated as Fe″, both in chlorite and in chloritoid:

$$\{(Fe, Mn)_{2.9}(Mg, Ca)_{1.8}Al_{1.3}\}^{vi}\{Al_{1.3}Si_{2.7}\}^{iv}O_{18}H_{7.9}$$

$$(Fe + Mn)/(Fe + Mn + Mg + Ca) = 0.62$$

(2) All Fe''' calculated as Al

$$\{(Fe, Mn)_{2.4}(Mg, Ca)_{2.3}(Al,Fe''')_{1.3}\}^{vi}\{Al_{1.3}Si_{2.7}\}^{iv}O_{18}H_8$$
$$(Fe'' + Mn)/(Fe'' + Mg + Mn + Ca) = 0.51$$
$$(Fe^{total} + Mn)/(Fe^{total} + Mg + Mn + Ca) = 0.56$$

and the calculated weight norms, for Fe''' as oxidized Fe'', are given in Table 2.

Other pertinent petrographic data on this rock follow.

Muscovite: d (001) = 9.98 A
Paragonite: d (001) = 9.62 Å
Chlorite: d (001) = 14.11 Å
β (Na light) = 1.642 ± 0.001

Rock no. 392-1. This rock is from a local 730′ prominence 0.6 mile S69°W of the 661′ benchmark near Butler Pond in Whipple Hollow, immediately west of the local road. It is mapped as the West Castleton(?)-Hortonville(?) formation (Zen, 1959b). In the hand specimen, it is a grey, massive rock with minute porphyroblasts of a feldspar studded throughout the sample. Thin section and x-ray analysis show the mineralogy to be albite, chlorite, muscovite, quartz, and rutile. On the muscovite, $d(001) = 10.00$ Å; β (Na light) $= 1.601 \pm 0.001$. The albite is untwinned and shows rounded outline; $\Delta 2\theta_{CuK}((\bar{2}01) - (10.0)_{Qtz}) = 1.17°$, near that for Amelia albite, $\Delta 2\theta = 1.18°$ and therefore presumably nearly pure $NaAlSi_3O_8$ since both calcium (Smith, 1956, p. 52) and potassium (Tuttle and Bowen, 1958, p. 13) tend to decrease this angular separation. This is confirmed by the value of $\alpha, < 1.540$, showing that it is substantially free of calcium.

The calculated mode, on the basis of pure potassic muscovite and no potassium in the feldspar, is given in Table 2; an Fe/(Fe+Mg) ratio of 0.64 is computed for the chlorite. The results tally with an x-ray mode, based on calibration curves and technique developed by the writer and described elsewhere (Schmalz and Zen, in preparation).

Rock no. 648-1. Specimen 648-1, from an outcrop of the Bomoseen greywacke at 1490′ elevation on a small terrace 1.6 mile N22°W of Biddie Knob, is a dark grey, massive mudstone showing occasional rock chips up to 3 mm. across and grains of quartz and feldspar as well. Thin section and x-ray studies show the following mineral assemblage: stilpnomelane, chlorite, muscovite, plagioclase, microcline, quartz. The stilpnomelane has these properties:

$$2\ V = 0°; \quad \beta\ (\text{Na light}) = \gamma = 1.687 \pm 0.002; \quad d\ (001) = 12.1\ \text{Å}.$$

Using the Michel-Lévy method, the albite turns out to be nearly calcium-free.

Because of the uncertain composition of both chlorite and stilpnomelane in this rock, it has not been feasible to calculate a norm for it.

REPRESENTATION OF MINERAL ASSEMBLAGES BY PHASE DIAGRAMS

Introduction. To study the mutual relations of the various minerals and mineral groups found in the slates, two possible approaches may be chosen. The first is to study each mineral assemblage as it is found, and attempt to see how it reflects the physical and chemical environment of formation; different assemblages may then be compared for the causes of the divergences. On the other hand, the mineral assemblages may also be studied by constructing multicomponent, isothermal, and isobaric phase diagrams, and correlating the observed associations with the various phase assemblages indicated by the diagrams. This second approach, though less direct, has many advantages. It emphasizes the compositional relationships among the minerals as well as the control of bulk composition on the mineralogy. It enables one to predict the changes in the mineralogy and compositions of individual phases with changing physical variables. Most significant, however, is its ability to bring together the different mineral assemblages, observed or as yet undiscovered, and exhibit their mutual relationships—which differences may reflect merely compositional variations, and which must reflect changing external factors such as temperature and pressure.

Because of these reasons, the second method is adopted in this study. A number of excellent general treatises of the method in the literature (Thompson, 1957; Korzhinskii, 1959) makes this feasible. Graphical methods will be used extensively. Because of the practical difficulty of representing multicomponent systems in two dimensions, where only three components can be conveniently depicted, standardized methods of selection of components are employed; the scheme is considered with each diagram. The diagrams to be considered are the following, in that order:

1. The A-N-K diagram: Al_2O_3; Na_2O; K_2O
2. The A-C-F(M) diagram: Al_2O_3; CaO; FeO+MgO
3. The A-K-F-M tetrahedron:
    a) The A-K-F(M) projection
    b) The A-F-M projection.

In the study of fine-grained rocks, it is not often possible to ascertain the exact compositions of all phases in a specimen, nor even always to identify all the phases present, so that one may be forced to use the bulk chemical analysis to estimate the composition of one phase when those of the other phases are known; indeed one may have to resort to this method to detect unsuspected phases. It is thus necessary to devise schemes for plotting a bulk chemical analysis onto any given ternary diagram. The particular scheme adopted depends of course on the nature of the other phases present and might change with each crossing of an isograd.

The A-N-K diagram. As no mineral found in the Castleton area has its gram-formula ratio of alkali to aluminum in excess of unity, corresponding to the alkali feldspars, the following values may be assigned to the apices of this ternary diagram: "A" = (Al_2O_3-K_2O-Na_2O); "N" = $NaAlO_2$; "K" = $KAlO_2$. The diagram, Fig. 3, is valid for quartz-bearing rocks and at some particular values of temperature, pressure, and the chemical potential of H_2O.

The proper aluminum silicate phase to place at the "A" corner, for the

Castleton area, is uncertain. As discussed before, Dale's report of pyrophyllite (1898, p. 191) from this area cannot be verified. The *x*-ray pattern of pyrophyllite is such that it should be easily detected even in small amounts. The writer suspects that in many slates kaolinite is the "A" phase. Kaolinite is difficult to ascertain in the presence of chlorite (Bradley, 1954), especially in small amounts, and could have escaped notice in many samples where it is expectable by the mineral association. Until more data exist, this phase is assumed to be appropriate for the "A" corner in the Castleton area; as we shall see, this does not affect any of the equilibrium arguments to be presented.

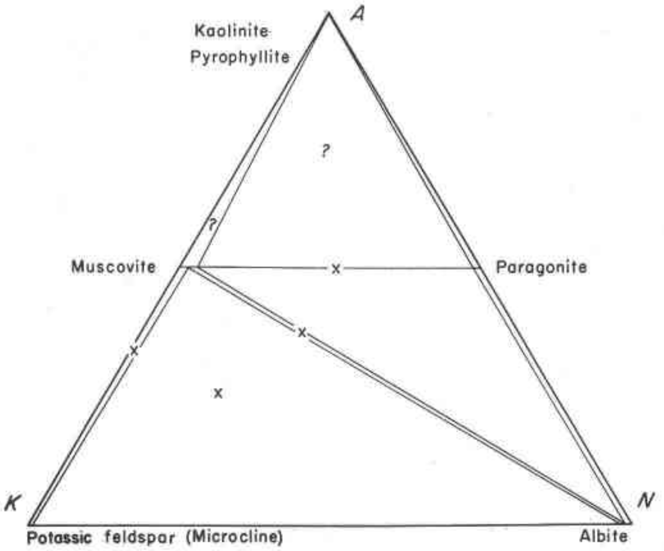

FIG. 3. The A-N-K diagram for rocks with free quartz and under particular values of the external conditions, appropriate to the Castleton region. Crosses indicate assemblages found in the Castleton area. If Dale's report of pyrophyllite is accepted, then one of the two assemblages marked with a question mark probably should be added. In this and in the ensuing diagrams, the position of all the phases are calculated according to the formula proportions of the oxide components.

In the Castleton area, the following mineral assemblages, appropriate to this diagram, have been found:

Muscovite-albite-potassic feldspar (at least in part microcline)
Muscovite-albite
Muscovite-paragonite

In addition, *x*-ray data on the slates of the Poultney River group indicate the muscovite-potassic feldspar assemblage, although due to interference by the muscovite pattern the presence of small amounts of albite in the samples cannot be disproved.

The muscovite-albite-potassic feldspar association defines a 3-phase field which under particular temperatures and pressures is invariant. The muscovite from the Bomoseen greywacke, which supplies this assemblage (at 1370′ level south of 1444′ knob near Walker Pond), has basal spacing of 10.00±0.02 Å and so is virtually pure $KAl_2AlSi_3O_{10}(OH)_2$. Unfortunately, no good *x*-ray method exists for the determination of the Na/K ratio of low-temperature potassic feldspar; its amount in a specimen is commonly so low or the size of the grains so fine that no reliable refractive index can be measured. It seems certain, however, that the feldspars must be nearly pure potassic and sodic phases, without appreciable solid solution.

The muscovite-paragonite relations have been considered in an earlier section. The muscovite in coexistence with paragonite in the Castleton area has a basal spacing of 9.975 Å; between this muscovite and those with 10.00 Å basal spacing, therefore, must lie all muscovites in the 2-phase muscovite-albite field, the 3-phase muscovite-albite-potassic feldspar field, and the 2-phase muscovite-potassic feldspar field. The determined spacings from these assemblages are given in Table 3. The data are consistent except those from specimens 100-3, 648-1, and 1219-1. The last specimen, from the Bascom formation at Halls Island near Brandon (Cady and Zen, in press), is a rock of considerably higher metamorphic grade and so its muscovite spacing may not be comparable with the rest. The low values for the other two samples, however, are not readily explained.

Many slates exist which do not have an essential sodic phase. This does not necessarily mean that the rock has no sodium in it; rather if the amount is sufficiently small it is easily camouflaged in muscovite.

Although in Fig. 3 paragonite is shown to be stable, the lowest-grade metamorphic rocks from which it is known are phyllites; in the Castleton area these rocks are confined to the Taconic Range. No paragonite has been found in the true slates despite systematic search. While possibly due to inadequate sampling, this fact could, in view of the coexistence of kaolinite and albite in sedimentary rocks (see, for instance, Coombs, 1954, p. 81), be due to a reaction of the sort:

$$\text{albite} + \text{kaolinite} = \text{paragonite} + \text{quartz} + H_2O$$

$$NaAlSi_3O_8 + Al_2Si_2O_5(OH)_4 = NaAl_2AlSi_3O_{10}(OH)_2 + 2SiO_2 + H_2O$$

where the second, balanced equation would apply to stoichiometric phases only. The reaction to the right generates free H_2O (either vapor or liquid water), and suggests a net entropy gain; paragonite is on the high-grade side of the reaction.[2] Whether this hypothesis is correct, or whether the recorded occurrence of sodic illite or brammellite (Bannister, 1943) in certain sedimentary rocks bespeaks to the contrary, only additional petrographic data can settle.

On the other hand, under similar conditions microcline and kaolinite are probably unstable together and muscovite should exist instead. Careful petrographic work by Sand (1956) shows that a potassic mica always intervenes between kaolinite and potassic feldspar in weathering, although Coombs (1954, p. 81) reports kaolinite and potassic feldspar, both as pseudomorphs after analcite in the same specimen, in the altered Triassic sediments in New Zealand.

TABLE 3. BASAL SPACINGS OF MUSCOVITE IN VARIOUS ASSEMBLAGES IN THE A-N-K DIAGRAM

Specimen No.	Locality and Formation (See Appendix I)	*d* (001) muscovite, ±0.005A	Assemblage
100-3	South end of Graham Hill (Bull formation)	9.96	Quartz, *muscovite*, chlorite, *albite*
111-2	Southeast side of Zion Hill, above the cliff (Bull formation)	9.98	Quartz, *muscovite*, chlorite, *albite*
126-1	Southeast side of Sargent Hill (West Castleton formation)	9.98	*Muscovite, albite*, chlorite, quartz
137-2	960′ WSW of Eagle Rock, east of Beebe Pond (Poultney River group)	9.98	*Muscovite, albite, potassic feldspar*, chlorite, dolomite, quartz
182-1	Cut on Route 30, near Crystal Ledge (West Castleton formation)	9.98	*Muscovite, albite*, quartz, chlorite (?)
212-2	2000′, about 1000 feet east of Ben Slide (Biddie Knob formation)	9.975	*Muscovite, paragonite* (*d* (001) =9.62(3)), chlorite, quartz, rutile
213-1	1850′ in small ravine, 1500 feet east of Ben Slide (Biddie Knob formation)	9.98	*Muscovite, paragonite* (*d*(001) =9.62) chloritoid, chlorite, quartz, rutile
392-1-2	730′ knob, 0.6 mile SWof Butler Pond in Whipple Hollow (West Castleton formation (?))	9.99	*Muscovite, albite*, chlorite, quartz, rutile
648-1	1490′, 1.6 miles NNW of Biddie Knob (Bomoseen greywacke)	9.95	*Muscovite, microcline, albite*, chlorite, stilpnomelane, quartz
722-1	1170′, north of Eagle Rock near Beebe Pond (Poultney River group)	9.97	*Muscovite, potassic feldspar*, quartz
982-1	650′, east of Scotch Hill Road, 0.6 mile north of 424′ BM along the road (Poultney River group)	9.98	*Muscovite, albite, potassic feldspar*, dolomite, chlorite, quartz
1219-1	East side of Halls Island near Brandon (Bascom formation)	9.95	*Muscovite, microcline, plagioclase*, chlorite, biotite (?), calcite, quartz
1747-1	600′, east side of ridge due east of 424′ BM, Scotch Hill Road (Poultney River group)	9.98	*Muscovite, potassic feldspar, albite*, chlorite, hematite, quartz

The A-C-F(M) diagram. The A-C-F(M) diagram has, for its apices, CaO, $FeO+MgO+MnO$, and (Al_2O_3-Na_2O-$3K_2O$). It is thus analogous to the original A-C-F diagram of Eskola (1915), and presupposes the

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[2] Although analcite and quartz are stable together under certain conditions (Coombs, 1954; Fenner, 1936; Milton and Eugster, 1959), analcite is not found in the Castleton area and will not be considered further.

presence of all K in muscovite and all Na in albite. Thus constructed, the diagram suffers the disadvantage, namely possible incompatible phases such as kaolinite and albite are not automatically excluded, where assemblages with potassic feldspar are not included. On the other hand, if the diagram is properly projected from albite such that paragonite occupies the "A" corner, phases like chloritoid may become "inaccessible" to the diagram and more than one projection, from different essential sodic phases, would be necessary. Such a procedure would be justified, however, only with more petrographic data than are now available. The pseudo-projection is therefore used with the reservations stated above. The diagram is shown in Fig. 4, in which the chemical potentials of CO_2 and H_2O are supposed to be such that calcite and dolomite, as well as the hydrous silicates, are stable. Unfortunately, our knowledge of the phases along the A-C boundary of the diagram is very poor. Calcite is certainly found in many samples; a mineral of the epidote group is found in one sample in the phyllites of the Taconic Range and zoisite in the marbles of West Rutland; but no other phase is known.

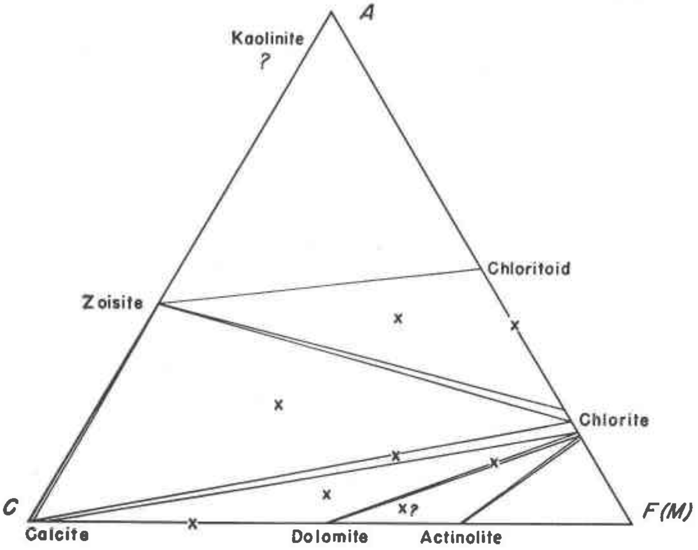

Fig. 4. The A-C-F(M) diagram for rocks with free quartz, under particular values of the external conditions appropriate to the Castleton area. Crosses indicate assemblages found in the Castleton area.

The mineral assemblages found in the Castleton area that belong to the diagram are as follows:

Chlorite-chloritoid-epidote[3]
Chlorite-calcite-dolomite
Chlorite-calcite
Dolomite-chlorite
Chlorite-chloritoid

In the adjacent marble belt of West Rutland, in addition, these associations are also found:

Calcite-chlorite-actinolite
Calcite-chlorite-zoisite

Since the "F(M)" corner is a superposition of at least two independent components, at least one more phase than shown on the diagram involving these components should be possible under arbitrary external conditions[4] without violating the Phase Rule. Unfortunately, *no* 4-phase assemblage belonging to this diagram has been found in the Castleton area and so the correct tie lines cannot be drawn.

In an earlier paper (Zen, 1959a) it was suggested that calcite and kaolinite could, under proper conditions, form a compatible pair in sediments and sedimentary rocks. The existence of zoisite rules out this pair, however, for rocks metamorphosed under conditions then prevailing in the West Rutland area. The formation of zoisite may be represented by the following reaction:

$$\text{calcite} + \text{kaolinite} = \text{zoisite} + CO_2 + H_2O$$
$$4CaCO_3 + 3Al_2Si_2O_5(OH)_4 = 2Ca_2Al_3Si_3O_{12}(OH) + 4CO_2 + 5H_2O,$$

the balanced equation being applicable if all the phases correspond to pure components. The reaction to the right entails a large increase in entropy due to the evolution of the volatiles, and (neglecting the volume of the volatiles) a volume decrease, so that this reaction is favored by increasing grade of metamorphism.

The assemblage calcite-dolomite-chlorite is a fairly common one in the Castleton area. From indices of refraction and x-ray data, the dolomite in coexistence with chlorite is sensibly free of iron, whereas the chlorite is relatively rich therein, as the following data for sample no. 102-2, from the south slope of Bull Hill, demonstrate:

Chlorite: $\beta = 1.630 \pm 0.003$ $\quad$ $Fe/(Fe + Mg) \approx 0.56$
Dolomite: $\omega = 1.862 \pm 0.002$.

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[3] Although this mineral may be ferriferous, the rock contains hematite and therefore this mineral belongs to the present diagram. The possible complication due to manganese is not considered.

[4] By "external conditions" is meant the collection of such independent variables as temperature, total pressure, and the chemical potentials of the mobile components. These are all intensive variables.

Actinolite is another phase belonging to the A-C-F(M) diagram. This mineral was reported by Dale (1898, p. 191) to occur extensively in the pelitic "Berkshire schist" of the northern Taconic Range; in the present study, however, chloritoid instead of actinolite is found in these areas, including those specifically mentioned by Dale. This is clearly a case of misidentification under the microscope. The only verified occurrence of actinolite is in the autochthonous Ordovician marbles of the West Rutland marble belt, coexisting with chlorite and calcite, and without dolomite.

Bowen (1940) predicted that for the metamorphism of a purely magnesian dolomite in the presence of quartz, tremolite is the first phase to form, and marks the disappearance of dolomite. Although this fits the observed association, dolomite is known to depart from the ideal formula of $CaMg(CO_3)_2$ both in the Ca/Mg ratio and in taking ferrous iron into solution. The need for dolomite to disappear at the tremolite-actinolite isograd, then, is no longer true and actinolite and dolomite of different Fe/Mg ratios may coexist. Indeed, in adjacent beds in the West Rutland area the quartz-dolomite-calcite assemblage is very common.

Another point to be remembered is that calcite itself could dissolve $MgCO_3$ and $FeCO_3$. Thus even if the rock is not sufficiently (Mg, Fe) rich to form the phase dolomite, actinolite might still form with progressive metamorphism by the reaction of quartz with calcite. This, however, will occur always at a higher metamorphic grade than the reaction involving dolomite, since the chemical potential of Mg in a calcite must be lower than in a dolomite under the same physical conditions. On the other hand, the occurrence, near the actinolite marble, of dolomitic marbles suggests that this is not the mechanism by which actinolite is formed in the West Rutland area, provided of course that the effects of possible variations in the bulk Fe/Mg ratios in these rocks are negligible. The need for caution in interpreting the actinolite isograd, however, is evident.

The assemblages (1) calcite-dolomite-chlorite; (2) calcite-kaolinite-chlorite; (3) calcite-dolomite-kaolinite; and (4) dolomite-chlorite-kaolinite have all been reported from unmetamorphosed sediments and sedimentary rocks (Zen, 1959a). In the A-C-F-M tetrahedron, in fact, all four phases may coexist, in the presence of quartz, under arbitrary values of the external conditions. For a particular Fe/Mg ratio, however, only three of these phases will in general obtain; for these four phases are related by a reaction of the sort

$$\text{calcite} + \text{chlorite} + CO_2 = \text{dolomite} + \text{kaolinite} + \text{quartz} + H_2O,$$

with more carbon dioxide released per unit reaction to the left than H_2O

consumed. We might thus expect the calcite-chlorite side to be relatively more stable at higher metamorphic grades. Rocks are found in the Castleton area which show the calcite-chlorite assemblage, with the chlorite, a high-Mg variety (optically positive and colorless), ramifying through the rock and apparently a result of metamorphic reaction between a carbonate and some pre-existing aluminum-rich phase (see the section on carbonate). It is possible that the above reaction is the correct one, although no dolomite is found in the rock today.

The A-K-F-M tetrahedron. For the A-K-F-M tetrahedron, the following coordinates have been used for the apices: A=(Al_2O_3-K_2O-Na_2O); K=$KAlO_2$; F=FeO+MnO; M=MgO. This construction then corrects for the presence of albite when a rock analysis is plotted; however, due to the possible incompatibility of albite and pyrophyllite (or kaolinite), the "A" corner should, in some instances, be paragonite; the difficulties are the same as has been discussed under the A-C-F(M) diagram.

In practice, this difficulty can be circumvented. The nature and relative amounts of the normative sodic phases can be estimated by plotting the "A" corner both as given above and as (Al_2O_3-$3Na_2O$), corresponding to paragonite, and noticing the relative proportions of these two versions at which the plot of the rock composition crosses the muscovite-chlorite join. This procedure is justified since below this join stilpnomelane, biotite, or microcline is a stable phase, none of which has been found with paragonite; and above this join chloritoid is stable, which so far is not found with albite in the Castleton area.[5]

In order to simplify the problem of graphical representation, two pseudoternary projections of the A-K-F-M system will be discussed.

1. A-K-F(M) diagram. This is the standard diagram suggested by Eskola (1915), wherein FeO and MgO are represented at the same apex. The phases that appear in this diagram, for rocks of the Castleton area, are shown in Fig. 5.* Among these phases, biotite and stilpnomelane alone show compositional variations *in* the pseudoternary system. However, biotite has not been found in enough samples and in sufficient quantities to permit a study of its phase relations in the Castleton area. Stilpnomelane, on the other hand, occurs abundantly and must be considered in some detail.

The projection of the stilpnomelane composition onto the A-K-F(M) diagram remains problematic. Although the alkali and alkaline earth

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[5] This procedure cannot be strictly correct since the muscovite-chlorite join is not a surface but a volume in the tetrahedron. This difference, however, will be overlooked.

* Note added in proof: Due to a drafting error, the chlorite-stilpomelane 2-phase field is shown to converge upon a fixed stilpomelane composition. In fact, stilpomelane in equilibrium with different chlorites should show different compositions also.

oxides in the chemical analyses of this mineral are dominantly potash, soda and lime do occur in variable amounts. In the section on the mineralogy of stilpnomelane it was seen that the best clustering of data results from assuming all the ferric iron as ferrous iron oxidized after the crystal had formed (although Shirozu's analysis clearly shows that this is not always true and some stilpnomelane has ferric iron playing the role of aluminum), and that generally the amount of calcium is so low that changing assumptions on its structural role affects the scatter of the plotted points but slightly. The hypothesis of total Fe as ferrous iron is therefore adopted, and on an A-K-F(M) projection all the analyses, whatever the role of Ca, loosely cluster below the biotite-chlorite join. Despite the higher SiO_2 content of stilpnomelane compared to chlorite and biotite, in the presence of quartz and albite, stilpnomelane may, to a good approximation, be taken as a phase on the A-K-F(M) diagram at approximately the indicated position.

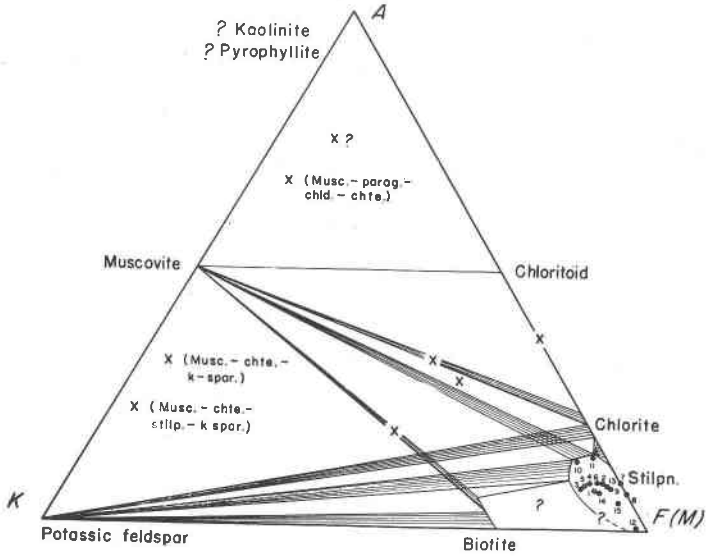

FIG. 5. The A-K-F(M) diagram for rocks with free quartz, under particular values of the external conditions, appropriate to the Castleton area. The numbers attached to the analyzed stilpnomelane have the same significance as in figure 1. Crosses indicate assemblages found in the Castleton area. Notice the apparent crossing of the tie-lines; this is due to the lumping of FeO and MgO as one component.

Since the "F(M)" corner is a superposition of two components, at least one additional phase (except along the A-K edge) may be expected in the diagram. Rocks from the Castleton area have yielded the following assemblages:

Muscovite-chlorite
Chlorite-chloritoid
Muscovite-chlorite-chloritoid
Muscovite-chlorite-potassic feldspar (in at least some cases definitely identified as microcline)
Muscovite-chlorite-stilpnomelane-potassic feldspar
Muscovite-biotite.

In addition, the assemblage muscovite-chlorite-dolomite-calcite is common among the two-carbonate marbles; when projected from calcite it properly belongs to the diagram. Since the assemblage muscovite-chlorite-potassic feldspar-stilpnomelane would exclude the association of muscovite and dolomite in the presence of calcite (provided that CO_2 does not act as an "inert" component), this might indicate that stilpnomelane does not truly belong to the diagram and calcium and/or ferric iron plays an essential role in its phase relations. That this is not always the case, however, is shown by the occurrence of hematite and calcite in a few stilpnomelane-bearing rocks; for these it must be reckoned as truly belonging to the A-K-F-M tetrahedron.

An alternative explanation is that stilpnomelane is stable with muscovite only at sufficiently high Fe/Mg ratios; at low ratios muscovite and dolomite become stable together instead. This idea is supported by the optically determined Fe/Mg ratio of stilpnomelane; in the Castleton area, also, the dolomite found with muscovite is always iron poor, judging by the index of refraction (see p. 155).

Mineral assemblages containing a potassic feldspar are rare in the Castleton area. These are found principally in the Bomoseen greywacke, associated with stilpnomelane; or in the Poultney River group and the Zion Hill quartzite, with muscovite, chlorite, and quartz. This scarcity of potassic feldspar is probably due to the fact that the bulk of the rocks have sufficiently high Al/alkali ratio so that muscovite is the stable potassic phase, rather than due to the intrinsic instability of potassic feldspar.

The muscovite-chlorite assemblage is a very common one in the slates of the Castleton area. Since the only other phases present in these rocks are quartz, albite, and rutile, four components, K_2O, Al_2O_3, FeO and MgO (ignoring any MnO that may be present) must be reckoned with. This suggests that for arbitrary values of the external conditions, the phases muscovite and chlorite must have a total of two independent means of compositional variation. One of these is accounted for by Fe-Mg substitution in chlorite; probably Al-Fe, Mg variation in chlorite accounts for the other degree of freedom.

As albite is known to coexist with muscovite and chlorite, without potassic feldspar, the chlorite apex, along the Al-Fe, Mg coordinate, of the albite-chlorite-paragonite 3-phase field must be at a higher Al/Fe, Mg ratio than the corresponding apex of the chlorite-muscovite-potassic feldspar 3-phase field.

2. The A-F-M diagram. The A-F-M diagram is constructed by projecting from muscovite which thus is assumed to be ubiquitously present (Thompson, 1957). The diagram thus constructed takes no account of calcic phases, and so is applicable only to calcium-free rocks, or to rocks in which no calcic phase involves the above components. The situation, where this is not valid, is complicated and data are very poor. Such rocks, luckily, are not important in the Castleton area.

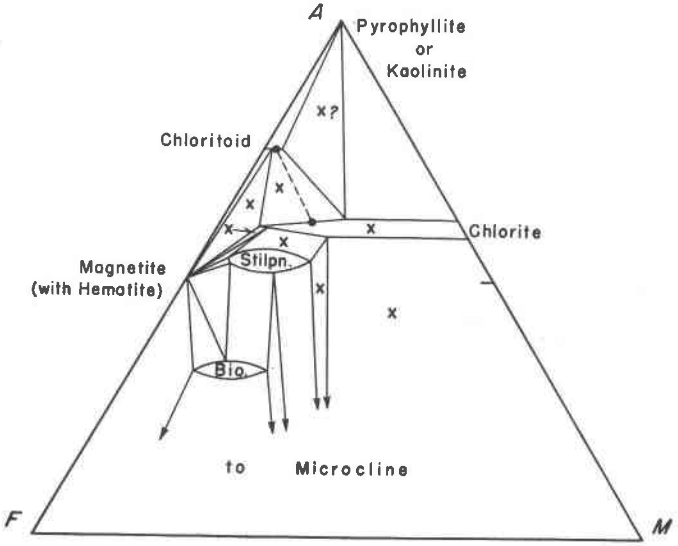

FIG. 6. The A-F-M diagram for rocks with free quartz and muscovite, under particular values of the external conditions, appropriate to the Castleton area. The diagram is a projection from muscovite composition in the A-K-F-M tetrahedron, according to the method of Thompson (1957). For rocks containing magnetite, an additional projection is made from hematite; this part of the diagram is thus strictly valid only where the rock contains both iron oxides. The lower portion of the diagram refers to rocks with negative values of "A"; it appears incomplete due to the method of projection. Crosses indicate assemblages found in the Castleton region. The dashed line from chlorite to chloritoid connects the analyzed chloritoid (no. 213-1A) with the calculated composition of the chlorite which is actually found in coexistence with the chloritoid. See text for details.

The diagram is given in Fig. 6, with quartz, muscovite, and a sodic phase to be added to those shown. Calcite or dolomite, but not both, may also be present. The lower side of the diagram is shown as unfinished; this is due to the geometry of the construction (Thompson, 1957). This fact does not, of course, preclude such assemblages as microcline-chlorite-stilpnomelane even on this projection; merely microcline is located at the unfinished lower boundary of the diagram.

The assemblages found in the Castleton area are listed below:

Chlorite-chloritoid
Chlorite-chloritoid-magnetite (ilmenite?), with hematite and rutile
Chlorite-stilphomelane
Chlorite-stilpnomelane-potassic feldspar (microcline in some samples)
Chlorite-potassic feldspar
Chlorite-magnetite
Chlorite
Biotite

If Dale's report of pyrophyllite from Ben Slide, in rocks now known to contain chloritoid and chlorite, is correct, another 3-phase assemblage would obtain.

It will be noticed that magnetite, and probably also stilpnomelane, contain ferric iron and therefore do not belong to this diagram. However, there exists stilpnomelane whose analysis shows virtually no ferric iron (Hutton, 1938, p. 184, 185); and as earlier discussed, the ferric iron in stilpnomelane may be due to later oxidation. As to magnetite, the one specimen that contains the indicated 3-phase assemblage also has hematite and rutile, and so even if the "magnetite" should turn out to be ilmenite (only a few specks exist and the amount was too small for specific identification), it still plots at the indicated position on the diagram. The diagram would then be a projection from rutile as well as from hematite.

On the other hand, large numbers of rocks from the Castleton area are without magnetite or indeed other phases appropriate to the "F" corner. Measured β of chlorites have not gone beyond 1.66, and with the exception of one specimen, beyond 1.64; the latter value corresponds to an Fe/(Fe+Mg) value of about 0.66. Thus the nature of the phase for the "F" corner of the diagram, in equilibrium with quartz and muscovite, remains in doubt. It might be iron-biotite, iron-chlorite, or an oxide; minnesotaite is unlikely with muscovite, as chlorite and potassic feldspar is a more probable alternative. Nor has any almandite been found in the Castleton area; this is presumably due to the low metamorphic grade.

The association of chlorite and chloritoid is a very common one in the Castleton region. One analysis of chloritoid was presented in Table 1; from this and the analysis of the rock the calculated approximate chlorite composition is given in Table 2. The values of β of chlorite from other, similar assemblages are given in Table 4, together with the estimated Fe/(Fe+Mg) ratios; the data indicate that chloritoid is always more ferroferrous than the coexisting chlorite.

TABLE 4. β-INDEX OF COEXISTENT CHLORITE AND CHLORITOID

Specimen number	β of chlorite, Na light, ±0.002	Est. Fe/(Fe+Mg) atomic[*]	β of chloritoid, Na light, ±0.002	Additional phases
6-3	1.630	0.56	1.718	Muscovite, quartz (G)
14-1	1.633	0.59	1.715	Muscovite, hematite, quartz (P)
19-1	1.632	0.58		P
24-1	1.635	0.61	1.718	G
25-2	1.634	0.59		P
36-1	1.634	0.59	1.72	G
43-1	1.640	0.66		G
55-1	1.630	0.56	1.722	G
213-1	1.642	0.67 0.62[‡]	1.722[†]	Muscovite, paragonite, quartz, rutile
217-1	1.628	0.54	1.72	Muscovite, paragonite, quartz, rutile
228-1-2	1.635	0.61		P
228-2	1.631	0.57		G
253-1	1.633	0.59	1.721	P
253-2	1.634	0.59		G
262-1	1.633	0.59		P
262-1	1.628	0.54	1.720	G
314-1-3	1.634	0.59		P
393-3	1.640	0.66	1.72	P
393-6	1.636	0.62	1.720	G
440-1	1.630	0.56		G
441-1	1.629	0.55		P
453-1	1.628	0.54	1.725	G
698-1	1.634	0.59		Muscovite, paragonite, quartz
726-1	2.638	0.64	1.720	G
726-2	>1.66[§]	>0.84	1.721	P
751-1	1.633	0.59		G

[*] Using Winchell and Winchell (1951, p. 385) and "$\frac{1}{3}$ substitution" formula.

[†] (Fe+Mn)/(Fe+Mn+Mg)≈0.92. See Table 1.

[‡] According to normative calculation from rock analysis, assuming the "$\frac{1}{3}$ substitution" formula. See Table 2.

[§] On one grain only. Data poor due to rarity of chlorite in the sample.

The chlorite-chloritoid 2-phase region is bounded on the two sides by two 3-phase regions. Unfortunately, no data on these 3-phase fields exist to set limits to the 2-phase region, nor the rate of movement of the limiting boundaries with metamorphic grade. The lowest grade rock containing the chlorite-chloritoid association (plus hematite, quartz, and muscovite), from near Lake Bomoseen, gives, on one grain of chlorite, $\beta > 1.66$, or an Fe/(Fe+Mg) ratio>0.84; this suggests that the 2-phase field may approach the pure iron end. At the high Mg side, a minimum estimate of the extension of the 2-phase field may be gotten by the index of refraction of the chlorite from this association, assuming that the lowest value of β for chlorite corresponds to the highest Mg content. The lowest value from table 4, 1.628 ± 0.002, corresponds to an Fe/(Fe+Mg) ratio of about 0.53, using the chart in Winchell and Winchell (1951) and assuming the "$\frac{1}{3}$ substitution" formula. The specimens, no. 262-1 and 453-1, respectively, came from 1630′ elevation on the southwest slope of Biddie Knob, and in the mountain brook at its crossing with Belgo Road, south of Grandpa Knob. Comparison with the index of 1.66 shows that the chlorite-chloritoid 2-phase field is fairly wide along the Fe-Mg coordinate for the Castleton area as a whole, although at any given locality this width may be considerably more restricted.

At the high Mg side of this 2-phase region, the limit being the 3-phase field chlorite-chloritoid-kaolinite or chlorite-chloritoid-pyrophyllite, the shift in the boundary with metamorphism may be given by the reactions

$$\text{kaolinite} + \text{chlorite} = \text{chloritoid} + \text{quartz} + H_2O$$

or

$$\text{pyrophyllite} + \text{chlorite} = \text{chloritoid} + \text{quartz} + H_2O.$$

The reactions to the right represent large entropy increases and net dehydration, as well as net volume decreases (ignoring the volatiles); thus they are favored by rising metamorphic grade. With rising grade, then, the chlorite-chloritoid-kaolinite (or pyrophyllite) 3-phase field will move to the high Mg side. This agrees with Hietanen's data (1951, p. 862, 867) from the Wissahickon schist, which is of higher metamorphic grade than the present slates. Hietanen's rock gives the mineral assemblages: albite-chlorite-chloritoid-muscovite, and the chloritoid analysis gives an Fe/(Fe+Mg) ratio of about 0.76. Assuming that all K occurs in muscovite and Na in albite, the normative chlorite calculated from the rock analysis gives for this same ratio a value of about 0.3.

The next 3-phase field shown on the diagram, chlorite-chloritoid-magnetite, cannot be discussed now because of insufficient data. Similarly, doubt exists as to the reality, in the Castleton region, of the 3-phase field magnetite-chlorite-stilpnomelane.

The association of chlorite and stilpnomelane is a fairly common one, albeit limited geologically to the Bomoseen greywacke. Accurate indices of refraction are difficult to obtain because of the fine grain size; correlation of indices of stilpnomelane with composition is uncertain due to lack of regression equations. One pair that has been measured, however, gives the following data:

Stilpnomelane: 2V(−) small; fast ray normal to cleavage.
Pleochroic, X = pale golden yellow
Y and Z = dark olive brown
$\beta = \gamma = 1.700 \pm 0.005$ (Na light)
Chlorite: $\beta = 1.648 \pm 0.003$.

Thus the stilpnomelane is probably more Fe rich than the coexisting chlorite. If Fe and Mg approximate ideal solution behaviors in both phases, this property may be expected to hold through the entire composition range over which the pair is stable.

Stilpnomelane-bearing rocks are of interest also in that these are among the few types in the Castleton area wherein a potassic feldspar occurs. Stilpnomelane, thus, might be regarded as the counterpart of chloritoid in low-Al rocks.

The assemblage of potassic feldspar and chlorite, with quartz and muscovite, is found in a few specimens of the Poultney River group argillites of Ordovician age, as well as in the Zion Hill quartzite. These, as well as those with stilpnomelane, are in fact the only instances of potassic feldspar in the area. From Fig. 6 it is seen that chlorite in coexistence with potassic feldspar may be expected to show the highest Fe, Mg-Al ratio in the presence of muscovite; however, chlorite-potassic feldspar assemblage without muscovite may give even higher ratios (see Fig. 5).

Many rocks in the Castleton area carry only muscovite, chlorite, and quartz, in addition to rutile, as detectable phases. As pointed out before, the frequency of such an assemblage suggests that it is not due to chance occurrence of proper bulk compositions but due to Al/(Fe+Mg) and Al/Si variabilities of such values as to remain in equilibrium with quartz (more formally,

$$\sum_k \frac{\partial \mu_{\mathrm{Si}}}{\partial \mathrm{N}_{k \neq \mathrm{Si}}} \Delta \mathrm{N}_k = 0$$

for a given set of external conditions, where μ_k is the chemical potential of component k and N_k its mole fraction), in addition to a variable Fe/(Fe+Mg) ratio in chlorite. On the A-F-M diagram, therefore, chlorite is shown as a region rather than a line; the variable Si content is not shown, but its chemical potential is fixed by that in quartz.

Difficulties are introduced in the treatment of the A-F-M diagram through uncertainties in the status of ferric iron. Since Fe''' and Fe'' may be related by an equation involving oxygen,

$$2\mathrm{FeO} + \tfrac{1}{2}\mathrm{O}_2 = \mathrm{Fe}_2\mathrm{O}_3,$$

they must be regarded as two independent components *if* oxygen were an inert component and is not counted as one itself. If, however, the chemical potential of oxygen is controlled by external conditions, it is mobile and Fe''' and Fe'' are to be regarded as a single independent component (Thompson, 1955, p. 81). Hematite, in this latter case, would plot on the A-F-M diagram and count toward 3-phase assemblages there.

If phases like magnetite and hematite coexist commonly in a given area, this fact would constitute powerful argument for regarding oxygen as inert, for an externally controlled value of μ_{O_2} would, in general, not correspond to the buffered μ_{O_2} value of the hematite-magnetite pair. On the other hand, to have only one of these phases, in particular hematite, could be due either to mobile oxygen or to excess oxygen.

In the Castleton area, hematite is the most common iron oxide phase and generally occurs without magnetite. Though suggestive of mobile oxygen, thus, the evidence is far from conclusive. Accordingly, indices of refraction of chlorite from the chloritoid-chlorite-hematite assemblage (purple) and *adjacent* (the distance varying, generally, from a few inches to a few feet) chloritoid-chlorite (green) assemblage were measured, as chlorite with hematite should be more Fe rich if hematite belongs to the A-F-M diagram, but not necessarily so otherwise. The results are given in Table 5. Although suggestive of a systematic variation with association, the data are not definite.

Recent studies of the iron formations of the Great Lakes region (James and Howland, 1955) show that for these rocks the mineralogy of the iron oxides indicates inertness of oxygen during metamorphism (Thompson, 1957, p. 855). Without further evidence to the contrary, this is the hypothesis tentatively adopted for rocks of the Castleton area.

Finally, the trace of the paragonite-albite boundary on the A-F-M diagram will be considered. Since both phases are known to coexist with chlorite, this trace must cross the chlorite field. In the presence of muscovite and quartz and under given external conditions, moreover, the system is univariant (four components and three phases) and so the trace may traverse the chlorite field in some arbitrary manner. Albite alone is found with stilpnomelane, and paragonite alone, to date, is found with chloritoid. Thus although the chlorite-stilpnomelane and chlorite-chloritoid 2-phase fields constitute limits between which the paragonite-albite boundary trace must lie in the Castleton area, nothing more definite can be said.

In the Wissahickon schist of Maryland, Hietanen (1941, p. 146; 1951, p. 862) reported the albite-chloritoid pair. The reaction

$$\text{paragonite} + \text{chlorite} + \text{quartz} = \text{albite} + \text{chloritoid} + H_2O$$

goes to the right with increasing metamorphic grade and so Hietanen's data from rocks of higher grade than those of Castleton are consistent with predictions. This also agrees with the higher Mg content in Hietanen's analyzed chloritoid. It should be noted that the system chloritoid-chlorite-paragonite-albite is invariant under specified external conditions and so the trace of the paragonite-albite field boundary must follow one of the chloritoid-chlorite tie lines. For this reason also, this boundary will always "jump" 3-phase fields on the A-F-M diagram. In general, with

TABLE 5. β-INDEX OF CHLORITE IN GREEN VS. PURPLE SLATES FROM THE SAME LOCALITY

Specimen number	Locality (See Appendix I)	Assemblage	β, ±0.002 (Na light)	Fe/(Fe+Mg)*
14-1	1650′ terrace, south of Biddie Knob	Chlorite, chloritoid, muscovite, hematite, quartz (P, purple)	1.633	0.59
217-1	1450′ in trail SSW of Biddie Knob, W of crest of Taconic Range	Chloritoid, chlorite, muscovite, paragonite, quartz, rutile (green)	1.628	0.54
262-1	1580′, WSW slope of Biddie Knob	Chloritoid, chlorite, muscovite, quartz, rutile (G, green)	1.628	0.54
262-1	Same specimen	P	1.633	0.59
6-3	1200′, pasture due west of Biddie Knob	G	1.630	0.56
253-2	1130′, same pasture	G	1.634	0.59
253-1	Same	P	1.633	0.59
228-1-2	990′, S. of Giddings Brook, just S. of 971′ road junction, N. of Sargent Hill	P	1.635	0.61
228-2	960′ in Giddings Brook, same locality	G	1.631	0.57
751-1	700′ in Giddings Brook	G	1.633	0.59
698-1	910′ in pasture S. of Giddings Brook, $\frac{1}{8}$ mile NE of 1010′ knob just S. of Monument Hill Road	G	1.634	0.59
726-1	540′, due W. of 800′ knob just southeast of the village of Hubbardton	G	1.638	0.64
726-1	Same locality	P	>1.66†	>0.84
440-1	Small knob north of Castleton River, $\frac{1}{2}$ mile SW of 1918′ knob, 5 miles east of Castleton	G	1.630	0.56
441-1	1405′ knob 200 feet south of last locality	P	1.629	0.55
393-6	1800′, west slide, Ben Slide	G	1.636	0.62
393-3	Same locality	P	1.640	0.66

* Using Winchell and Winchell (1951, p. 385) and "$\frac{1}{3}$ substitution" formula.

† On one grain only. Data poor due to rarity of chlorite in the sample.

increase in metamorphic grade the trace of the paragonite-albite boundary on the A-F-M diagram moves towards the "A" corner, and the field of albite expands at the expense of the field of paragonite.

EVIDENCE FOR PHASE EQUILIBRIUM

The foregoing discussions of mineral reactions in the rocks of the Castleton areas, based on phase diagrams derived ultimately on thermodynamic principles, assume chemical equilibrium among the phases under the physical and chemical conditions of metamorphism. It is necessary, therefore, to present evidence that such equilibrium is indeed attained in the rocks.

One of the most powerful criteria for chemical equilibrium in any system is its obedience to Gibbs' Phase Rule. Obedience of this Rule is a necessary, though not a sufficient, condition for equilibrium. The test consists simply of counting the number of components and phases, such that

$$\phi \leq c + 2$$

where ϕ is the number of phases and c the number of independent components; if mobile components obtain, of which there are c' in number,

$$\phi' \leq c + 2 - c'$$

(Thompson, 1955, p. 81). For arbitrary values of temperature and hydrostatic pressure, a condition applicable to a randomly chosen sample, we have

$$\phi' \leq c - c'.$$

In rocks of the Castleton region, even assuming H_2O and CO_2 to be mobile components, this restricted version of the Phase Rule is never violated.

Even so, however, a rock may be out of chemical equilibrium if phases, empirically determined to be mutually incompatible, are found together. Such incompatibility would be demonstrated if, for example, two phases striding the composition of some intermediate phase, known to be stable, are found together. In the presence of quartz, this might be, for instance, microcline and pyrophyllite instead of microcline-muscovite or muscovite-pyrophyllite. Such incompatible phases are not found together in the Castleton area.

In many purple and red slates, there are thin beds of pyritic quartzite or of limestone, as well as pebbles of limestone in a slaty matrix. In all instances, except where the limestone itself is salmon-colored and carries hematite, there is a layer, though possibly an extremely thin one, of green slate between the two lithologies. This relation is so consistent as to suggest some incompatible relation between some phases of the two rock types. The nearly ubiquitous presence of rims like this implies the existence of chemical reaction among the minerals, and at least locally the attainment of equilibrium.

The argument for phase equilibrium would be more persuasive if it can be demonstrated that the phase relations depicted in the previous diagrams are obeyed *in detail.* For example, consider the relative Fe/(Fe+Mg) ratio in the chlorite-chloritoid 2-phase field vs. the Na/(Na+K) ratio in the albite-muscovite-microcline 3-phase field. If equilibrium obtains, adjacent rocks of presumably identical metamorphic grade should show variation in the ratio of the 2-phase assemblage corresponding to bulk compositional variations, but invariant ratios for phases in the 3-phase assemblage. At the same time, adjacent beds should show different phase assemblages corresponding to appropriate bulk compositional differences. Unfortunately, lack of sufficiently coarse mineral grains suitable for chemical analysis greatly hampers the use of the test. Moreover, refractive indices of chlorite and especially of chloritoid, even if correlated directly with Fe/(Fe+Mg) ratios, show rather narrow ranges of variation and cannot be used confidently to check, for instance, if the 2-phase compositional tie lines "cross" from one sample to another, thus suggesting possible disequilibrium. Finally, there are relatively few 3-phase assemblages in the rocks for such a study.

On the other hand, there is no evidence to refute the idea of equilibrium, and the mineral associations observed do "make sense." Thus one of the few associations of potassic feldspar found, in the Bomoseen greywacke, is with stilpnomelane and albite—where (for example, near Brandon Mountain Road) nearby on *both sides* of this rock unit the Biddie Knob formation carries no feldspar but does carry chloritoid and locally paragonite. This is clearly a case of mineralogical variation due to bulk composition; such consistent results are not to be expected from chance collection of detrital material.

The compositional arguments of phase equilibrium may be supplemented by textural data. Clearly, porphyroblasts in a rock demonstrate recrystallization during metamorphism and thus probably in equilibrium with its surroundings during some stage. Feldspar, mainly untwinned albite, and chloritoid porphyroblasts are abundant in the Castleton area. Recrystallized limestone and dolomite, as well as minerals in fracture and boudinage fillings, again testify to mineralogical and textural response to metamorphism and involve chemical reconstitution.

The platy minerals also show definite evidence of recrystallization. Microporphyroblasts of chlorite and of muscovite, or intergrowth of these two, on the order of 0.1 mm. across, or of either of these with chloritoid, are extremely common in the slate. These crystals are randomly oriented but generally cut across the slaty cleavage as well as lying within the cleavage surface. The orientations are also random with respect to the bedding, where this is visible. Compositionally, there is no measurable difference between the porphyroblastic chlorite and chlorite in the groundmass and suggests that all the crystals are recrystallized, the difference in size being controlled by growth rate or surface tension.

Stilpnomelane in the Bomoseen greywacke tends to be extremely fine-grained, on the order of 20μ; these are thin books which commonly wrap around larger grains of quartz and feldspar, as well as occurring in sheaves. Such textures are unlikely due to primary sedimentation and/or deformation alone and probably result from recrystallization.

Some phyllites from the Taconic Range show segregation of the platy minerals and quartz into distinct bands on a microscopic scale. The platy minerals show parallel orientation with the basal planes in the plane of foliation. The quartz grains show roughly rectangular outlines in thin sections, with two edges parallel to the foliation. Such a geometric relation cannot be altogether primary and must be due to recrystallization with approach to textural equilibrium. Chemical readjustment clearly also took place, with chances of attainment of equilibrium.

Summary and Conclusions

Perhaps the single most important conclusion to be drawn from the present investigation is the fact that rocks at the lower end of the metamorphic spectrum are amenable to phase-equilibrium studies. With the aid of modern *x*-ray techniques, the mineralogy of even the slates can generally be ascertained to a fair degree; the grain sizes are not so small as to bring in uncertainties due to poor crystal quality, as is commonly the case with sediments. What data the *x*-ray method fails to furnish can commonly, though not always, be furnished by microscopic examinations; among these are textural information on the rocks.

Not only is the mineralogy of a rock determinable, but the results can be analyzed in terms of phase diagrams. The relation among the various minerals and mineral assemblages can be understood and can be compared directly with similar results from rocks of higher metamorphic grades. Despite the fact that an observed mineral assemblage generally comprises too many independent components to make possible a complete graphical representation of the phase relations, it has been found possible, by and large, to eliminate enough of them by proper projections, so that only three or four of these appear in any one diagram and so can be graphically shown. The limitations imposed by these simplifying assumptions do not vitiate their usefulness.

The second conclusion that might be drawn concerns the existence and nature of isograds in the low-grade argillites. Chloritoid seems to be one of the best isograd-indicators; indeed it was used by Balk (1953) south of the present area. In the present area, this mineral is confined to east of Lake Bomoseen, and is most widespread in the lustrous, relatively coarse-grained phyllites of the Taconic Range. However, the fact that this mineral is not found in *all* pelites east of its first appearance, but is confined to a stratigraphically mappable unit, bespeaks the importance of the bulk composition in its existence, for a given set of conditions. Thus its non-existence in the western part of the area might be due as much to rock composition as to lack of the proper physical environment.

Stilpnomelane has been found throughout the area regardless of the metamorphic grade; if a lower isograd exists for it, this must occur in rocks of even lower grade. For the same reason, little can be said at the moment of the biotite isograd.

Chlorite is found throughout the area and no isograd exists for this phase. Recent studies of sediments and sedimentary rocks show that this is one of the commonest ferromagnesian phases and it is questionable whether chlorite, as a phase, has any lower stability limit at all within the accessible part of the earth's surface.

A third general conclusion that might be drawn is the essential simplicity of the mineral assemblages. With few exceptions, the minerals are well-known, and they occur in combinations each of which includes a small number of phases. This indicates, first, that the systems are in mutual equilibrium, for it is not likely that a chance collection of detrital grains will give such ordered simplicity; second, that many of the phases must have solid solubility along one or more compositional coordinates, to rather extensive degrees. This is in fact borne out by the mineralogical studies. Finally, the mineralogical compositions of the rocks are quite consistent with those of higher metamorphic grades and there is no evidence for breaks in the nature of the metamorphism at lower grades nor any unusual trends.

A fact of some interest is that chloritoid and stilpnomelane are found in the same metamorphic belt. Both minerals have been commonly reported from areas of low grade metamorphism—however, as a rule, from different areas. In the Castleton area, the stilpnomelane-bearing Bomoseen greywacke occurs not only side by side, but repeatedly interfolded, with the chloritoid-bearing Biddie Knob formation (Zen, 1959*b*). The physical conditions of metamorphism could not have been very different, and the existence of these two phases must be controlled by chemical compositions of the rocks. As has been indicated earlier, this is borne out by their respective mineral assemblages.

The discussions also make clear the fact that our information on the low-grade metamorphism of argillaceous sediments is far from adequate; much more data are needed, from this general area as well as from other regions, before many questions can be answered. For example, many mineral assemblages have been predicted by the diagrams but are not yet observed. The projection of the paragonite-albite transition on the A-F-M diagram is only partly defined. The phase occupying the "A" corner of the diagrams in the presence of quartz is not certain. The role of stilpnomelane on the A-F-M diagram, if it belongs there at all, is far from fully understood. We need more information to check if paragonite has a lower limit of stability, giving way to kaolinite and albite. The whole problem of the A-F-M projection for calcareous rocks has been sidestepped for lack of data. Few of the "3-phase assemblages" in the pseudoternary projections have had their apices pinned down and their rates of change with progressive metamorphism delimited. Finally, the part played by redox reactions in the mineral assemblages remains obscure, *i.e.*, we do not know for sure what is the best way to treat oxygen as a component, whether as a mobile component, and therefore treating ferric phases as if they were ferrous phases in the A-F-M diagram, or as an inert component, so that ferric and ferrous iron must be regarded as independent components. Although present indications are in favor of the latter, the situation is by no means unequivocal.

Appendix I. Location of Geographical Names Mentioned in the Text

(All locations given to the nearest 1′ of latitude and longitude. Names in parentheses refer to names of U. S. Geological Survey topographic sheets.)

Name	*Longitude* (73° W+)	*Latitude* (43° N+)
Beebe Pond (Bomoseen)	11′	44′
Belgo Road (Bomoseen; Proctor)	06′–08′	38′
Ben Slide (north side of 2390′ knob) (Castleton)	06′	34′
Benson (Benson)	19′	42′
Bird Mountain (Castleton)	07′	35′
Brandon (Brandon)	06′	48′
Brandon Mountain Road (Bomoseen)	08′	44′
Biddie Knob (Proctor)	06′	43′
Bull Hill (Bomoseen)	11′	38′
Butler Pond (Proctor)	05′	42′
Castleton (Castleton)	10′	37′
Castleton River (Castleton; Thorn Hill)	04′–17′	36′
Clark Hill (Castleton)	04′	34′
Crystal Ledge (780′ knob) (Bomoseen)	11′	39′
Eagle Rock (Bomoseen)	11′	44′
Giddings Brook (Bomoseen)	08′–11′	42′–43′
Glen Lake (Bomoseen)	14′	40′
Graham Hill (Castleton)	11′	37′
Grandpa Knob (Proctor)	06′	40′
Gully Brook (Castleton)	08′	36′
Halls Island (Brandon; Sudbury)	07′	47′
Herrick Mountain (Castleton)	06′	33′
Hubbardton (Bomoseen)	11′	42′
Lake Bomoseen (Bomoseen; Castleton)	11′–14′	36′–42′
Mill Pond (Benson)	17′	42′
Monument Hill Road (Bomoseen)	10′	43′
Sargent Hill (Bomoseen)	09′	43′

Scotch Hill Road, 424′ Benchmark (Bomoseen)	15′	38′
Taconic Range (Castleton; Bomoseen; Proctor; Sudbury)	05′–10′	37′–47′
Walker Pond (Bomoseen)	08′	44′
West Rutland (Castleton)	03′	36′
Whipple Hollow (Proctor)	04′–05′	36′–42′
Zion Hill (Bomoseen)	09′	41′
1405′ knob north of Castleton River (Castleton)	07′	37′
1930′ knob west of Clark Hill (Castleton)	05′	35′
2035′ knob (Proctor)	07′	41′

Appendix II. Location of Specimens Mentioned in the Text

(Names in parentheses refer to names of the formations)

6-3	1100′ in pasture 1.0 mile northwest of Biddie Knob, near 1007′ benchmark (green; Biddie Knob)
14-1	1650′ on terrace 0.3 mile due south of Biddie Knob (purple; Biddie Knob)
19-1	1300′, east slope of 1630′ peak at the north end of Bird Mountain (purple; Biddie Knob)
24-1	1750′, north of topographic saddle, 1.2 miles north-northeast of Herrick Mountain (green; Biddie Knob)
25-2	760′ in major north-flowing brook east of Bird Mountain (purple; Biddie Knob)
29-4-1	1400′ on the west slope of Clark Hill (purple; Biddie Knob)
36-1	1630′ knob 0.6 mile east of Grandpa Knob (green; Biddie Knob)
43-1	620′ in Gully Brook, at junction with tributary from the west (green; Biddie Knob)
55-1	1240′ on ridge, 0.6 mile northwest of 635′ road corner 1.3 miles northeast of Biddie Knob (green; Biddie Knob)
100-3	480′, 0.2 mile south-southwest of Graham Hill (green; Mettawee slate, Bull)
102-2	600′, 0.7 mile south-southwest of 905′ top of Bull Hill (green reaction rim between purple Mettawee slate and dolomitic nodule)
111-2	1010′ knob due west of 816′ benchmark east of Zion Hill (green; Mettawee slate, Bull)
126-1	1250′, about 0.2 mile south of 1360′ knob of Sargent Hill (West Castleton)
137-2	960′ west-southwest of Eagle Rock, east of Beebe Pond (black slate; Poultney River group)
182-1	Cut on Route 30, west of Crystal Ledge (West Castleton)
197-2	720′, 0.2 mile south-southeast of top of Crystal Ledge (green; Mettawee slate, Bull)
212-2	2000′, 1000′ east of Ben Slide, S75°W from 1930′ knob west of Clark Hill (green; Biddie Knob)
213-1	1850′, in northeast-trending ravine 1500′ east of Ben Slide (green; Biddie Knob)
228-1	990′, south of Giddings Brook, just south of 971′ road junction, north of Sargent Hill (purple; Biddie Knob)
228-2	960′ in Giddings Brook, same locality (green; Biddie Knob)
233-1-2	1270′, 0.2 mile southeast of 1360′ knob of Sargent Hill (green; Mettawee slate, Bull)
253-1	1130′, in pasture 0.9 mile northwest of Biddie Knob, near 1007′ benchmark (purple; Biddie Knob)
253-2	same locality (green; Biddie Knob)
262-1	1580′ on west-southwest slope of Biddie Knob (purple and green; Biddie Knob)

314-1	1990′ knob in Taconic Range, 0.5 mile south-southeast of 2035′ knob (purple) Biddie Knob)
392-1	730′ prominence, 0.6 mile S69°W of 661′ benchmark near Butler Pond (West Castleton?)
393-3	1800′, west slide, Ben Slide (purple; Biddie Knob)
393-6	same locality (green; Biddie Knob)
440-1	small knob 200′ north of 1405′ knob north of Castleton River (green; Biddie Knob)
441-1	1405′ knob north of Castleton River (purple; Biddie Knob)
453-1	860′ in brook just south of Belgo Road (green; Biddie Knob?)
648-1	1490′ on small terrace 1.6 mile N22°W of Biddie Knob (Bomoseen)
698-1	910′ in pasture south of Giddings Brook, $\frac{1}{8}$ mile northeast of 1010′ sharp knob just south of Monument Hill Road (green; Biddie Knob)
722-1	1170′, north of Eagle Rock near Beebe Pond (black slate; Poultney River group)
726-1	540′ due west and on slope of 800′ knob just southeast of the village of Hubbardton (green; Biddie Knob)
726-2	same locality (purple; Biddie Knob)
751-1	700′ in Giddings Brook (green; Biddie Knob)
982-1	650′, east of Scotch Hill Road 0.6 mile north of 424′ benchmark along the road (green slate; Poultney River group)
1219-1	470′ knob at east end of Halls Island (Bascom)
1747-1	600′, east side of ridge, due east of 424′ benchmark, Scotch Hill Road (purple slate; Poultney River group)

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