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# Coexisting paragonite and quartz in sillimanitic rocks from New Mexico

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## Abstract

Paragonite has been identified by electron microprobe and X-ray diffraction analysis in four specimens of metamorphic rock from northern New Mexico. The sodic mica coexists with quartz in three of these rocks. One comes from an area where kyanite, andalusite and sillimanite coexist. The other two come from widely separated areas where sillimanite is the only polymorph of $Al_2SiO_5$ to be found, and sillimanite coexists with paragonite and quartz in each. The paragonite–quartz–sillimanite assemblages appear to be stable. These are the first reported occurrences of this assemblage, and its presence supports the $Al_2SiO_5$ triple point of Holdaway (1971).

Muscovite occurs in two of the paragonite–sillimanite quartzites. Because these two samples crystallized at similar temperatures, and define a solvus, K/(K+Na) ratios should be similar for the two muscovites and for the two paragonites. However, large differences in K/(K+Na) of muscovite exist between the two samples and smaller differences in that ratio occur between the two paragonites. These variations may be related to differences in the celadonite content of muscovite in the two rocks. Consistent with tentative suggestions of previous workers, data presented here suggest that, as Fe and Mg are added to muscovite–paragonite pairs, K/(K+Na) increases in muscovite and may decrease slightly in paragonite. This relationship has serious implications for muscovite–paragonite solvus geothermometry since most natural muscovites and paragonites have small amounts of Fe and Mg. Generally, muscovite–paragonite solvus geothermometry seems to yield inconsistent and unreasonably high temperatures.

## Introduction

No assemblages with coexisting paragonite and quartz are known from sillimanite-grade metamorphic rocks (Chatterjee, 1972), but paragonite–quartz assemblages are relatively common in rocks containing kyanite or andalusite (Zen and Albee, 1964; Guidotti, 1968; Thompson *et al.*, 1977; Rumble, 1978; Hoffer, 1978; Labotka, 1980; Mohr and Newton, 1983). Chatterjee (1972) has interpreted this to mean that the assemblage paragonite + quartz breaks down to albite + $Al_2SiO_5$ at a temperature below the Al-silicate invariant point. According to experimental studies of paragonite + quartz breakdown, this would require the $Al_2SiO_5$ invariant point to occur at a temperature above 550°C (Greenwood, 1976). Such a temperature would support the experimental determination of the kyanite–andalusite–sillimanite equilibrium by Richardson *et al.* (1969).

Recent field and thermodynamic studies have suggested that the lower-temperature $Al_2SiO_5$ invariant point proposed by Holdaway (1971) may be more applicable to natural systems than that of Richardson *et al.* (Navrotsky *et al.*, 1973; Anderson *et al.*, 1977; Carmichael, 1978; Holdaway, 1978; Grambling, 1981, 1982; Hodges and Spear, 1982; Kieffer, 1982). A restricted stability field for paragonite–quartz–sillimanite should exist in nature if Holdaway's phase diagram for $Al_2SiO_5$ is correct, because Holdaway's sillimanite field overlaps Chatterjee's paragonite + quartz field between 3 and 5.5 kbar, 500–600°C. The conspicuous absence of paragonite–quartz–sillimanite assemblages from the published literature is problematic for supporters of Holdaway's (1971) $Al_2SiO_5$ invariant point.

If the assemblage paragonite–quartz–sillimanite exists in nature, the Precambrian terrane of north-central New Mexico should be an ideal place to find it. Rocks which contain coexisting kyanite, andalusite and sillimanite are widespread, petrologic evidence suggests that the coexisting Al-silicates crystallized close to equilibrium, and common mineral assemblages indicate metamorphic conditions of 500–550°C, 3.7–4.7 kbar (Holdaway, 1978; Grambling, 1981, 1982).

The present study documents the existence of four samples which contain paragonite from this area. All samples occur in sillimanite-grade rocks, three samples

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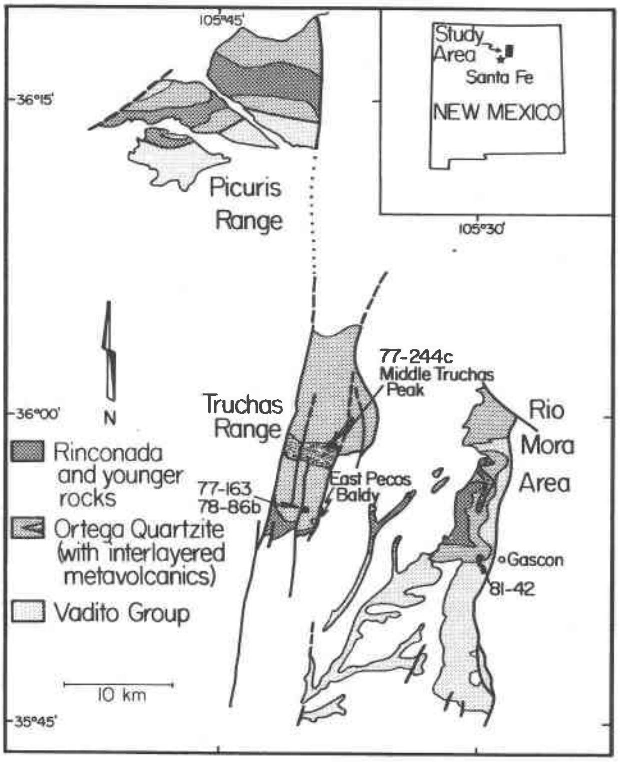

Fig. 1. Geologic map of the southern Sangre de Cristo Mountains, New Mexico, showing sample localities. Geology taken from Montgomery (1953), Miller *et al.* (1963), Gresens and Stensrud (1974), Grambling (1981 and unpublished data), Robertson and Moench (1979), Grambling and Codding (1982), Holcombe and Callender (1982), and Grambling, Williams and Codding (1983).

contain quartz, and two samples contain coexisting paragonite–quartz–sillimanite.

## *Regional geology*

The Picuris, Truchas and Rio Mora Ranges form three fault-bounded uplifts in the southern Sangre de Cristo Mountains, New Mexico (Fig. 1). Each uplift consists of a core of Precambrian igneous and metamorphic rocks overlain locally by a thin veneer of unmetamorphosed Paleozoic cover. Precambrian metamorphic lithologies can be divided into two stratigraphic groups, the Vadito and Ortega (Montgomery, 1953; Miller *et al.*, 1963; P. E. Long, 1976; Grambling and Codding, 1982). The older Vadito Group includes amphibolite, metarhyolite, meta-arkose and schist. The younger Ortega Group consists of massive, crossbedded orthoquartzite (the Ortega Quartzite) overlain by a sequence of pelitic schist, schistose quartzite and graphitic schist.

These stratigraphic units have been deformed by two episodes of folding, with local overprinting by third-generation folds (Nielsen, 1972; Grambling and Codding, 1982; Holcombe and Callender, 1982). $F_1$ folds are isoclinal, recumbent structures with a strong axial-plane schistosity and local transposition. $F_2$ folds are upright and tight to isoclinal with easterly strike and steeply south-dipping axial planes. $F_3$ folds are localized, gentle, open, upright structures with northerly trend, best developed in the Picuris and Truchas Ranges.

Peak metamorphic conditions were attained after $F_2$ folding as documented by porphyroblasts which cut $S_2$ cleavage and by isograds which cut axial planes of $F_2$ folds (Grambling, 1981). Metamorphic grade is characterized by the widespread coexistence of kyanite, andalusite and sillimanite. The three polymorphs occur in the Picuris Range together with either chloritoid or staurolite, and Holdaway (1978) inferred conditions of 530°C, 3.7 kbar for their coexistence. Grambling (1981) recognized kyanite, andalusite and sillimanite coexisting in part of the Truchas Range and calculated *P* and *T* as 535°C, 4 kbar based on the assemblages chloritoid–staurolite–$Al_2SiO_5$, cordierite–biotite–chlorite–muscovite–$Al_2SiO_5$, and on garnet–biotite geothermometry. The three Al-silicates coexist only in the central part of the Truchas Range. Elsewhere rocks contain a systematic distribution of one or two polymorphs and can be divided into a kyanite zone, a sillimanite zone and a zone where kyanite and andalusite coexist (Fig. 2). Three coexisting Al-silicates

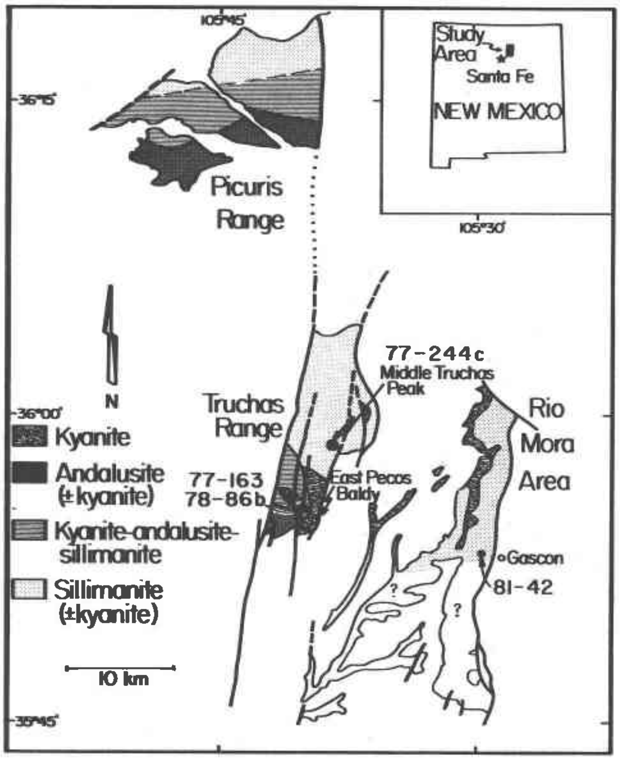

Fig. 2. Map showing the distribution of $Al_2SiO_5$ minerals in the southern Sangre de Cristo Mountains. Data from the Truchas Range and Rio Mora area are from Grambling (1981, 1982 and unpublished data). Data from the Picuris Range are only approximate, taken from Montgomery (1953), Holdaway (1978), McCarty (1983) and Grambling (unpublished data).

Fig. 3. X-ray photograph showing the distribution of K in intergrown muscovite and paragonite, sample 81–42. Scale bar is 10 μm long.

also appear in Rio Mora. Kyanite and sillimanite are widespread, but crystals of andalusite occur only along a single bed (the Vadito–Ortega contact) where they are enriched in $Fe_2O_3$ and $Mn_2O_3$ which have stabilized the andalusite to abnormally high pressure and temperature. *P–T* conditions at Rio Mora were near 550°C, 4.7 kbar during metamorphism (Grambling, 1982).

The age of metamorphism across the region is not well-documented. However, similar rocks in nearby areas have yielded apparent metamorphic ages near 1425 m.y. (L. E. Long, 1972; Gresens, 1975).

## *Methods*

This study is based on 56 specimens, four from the Picuris Range and the rest divided evenly between the Truchas and Rio Mora uplifts. Each specimen contains white mica. Micas were analyzed on an automated ARL-EMX microprobe at the University of New Mexico in a systematic search for paragonite. Operating conditions were 15 kV, 0.005–0.05 μA sample current on brass, 10–30 seconds counting time and a beam diameter of 1–3 μm. Qualitative X-ray elemental photographs were taken at the same conditions on the JEOL Superprobe 733 in the Department of Geology. Data were reduced using a Bence–Albee (1968) correction scheme, with natural and synthetic silicate and oxide minerals as standards. Error estimates are ±3 percent of the amounts present for most elements, ±10 relative percent for BaO and ±50 relative percent for F. The large errors on F analyses reflect the fact that F occurs near the limits of detection in most samples.

Micas were analyzed in three samples using X-ray diffraction techniques. Mica grains were hand-picked from the samples, crushed, mounted on glass fibers and analyzed in a Debye–Scherrer X-ray camera.

## Mineralogy

Of the 56 mica samples analyzed by electron microprobe, only four contained paragonite. Localities of these four samples are shown in Figures 1 and 2. All paragonite-bearing samples were collected from the lowest 250 m of the 1000 m-thick Ortega Quartzite.

Pure paragonite separates were obtained from two specimens, 77-163 and 77-244c. X-ray diffraction analysis has confirmed that the sodic mica is paragonite. Sample 77-163 contains the 1M polymorph with $d_{001}$ of 9.69Å, whereas sample 77-244c contains the more common $2M_1$ polymorph with $d_{002}$ of 9.68Å. White mica in sample 81-42 was analyzed but yielded a diffuse X-ray pattern. The diffuse pattern probably represents the fine intergrowth of muscovite and paragonite that can be seen in scanning X-ray photographs of this sample (Fig. 3).

Microprobe analyses of paragonite (Table 1) show that its composition varies considerably. One specimen (78-86b) has nearly pure $NaAl_3Si_3O_{10}(OH)_2$, with traces of FeO and F as the only impurities. Paragonite is less pure in other samples, with the most abundant impurity being K which replaces up to 14 mole% of the Na. Calcium is an important substituent in specimen 77-244c where it occupies 11 mole% of the alkali site. This sample has excess

Table 1. Microprobe analyses of white micas from paragonitic rocks. Note that paragonite in 78-86b occurs only as inclusions in garnet and does not coexist with muscovite or quartz.

| | 77-163 P | 77-244c P | 77-244c M | 78-86b $P^2$ | 78-86b M | 81-42 P | 81-42 M |
|---|---|---|---|---|---|---|---|
| FeO | .63 | .62 | 2.99 | 1.11 | 1.64 | .71 | 1.40 |
| MgO | .06 | .04 | .91 | .04 | .43 | .03 | .20 |
| MnO | .00 | .01 | .01 | .00 | .02 | .00 | .00 |
| $TiO_2$ | .08 | .05 | .28 | .02 | .22 | .08 | .28 |
| $Al_2O_3$ | 39.75 | 41.55 | 35.19 | 40.43 | 36.88 | 40.47 | 37.29 |
| $SiO_2$ | 45.20 | 44.53 | 44.77 | 47.73 | 46.34 | 45.20 | 45.12 |
| $K_2O$ | .96 | 1.18 | 9.42 | .06 | 7.97 | 1.41 | 7.28 |
| $Na_2O$ | 6.88 | 6.30 | 1.62 | 7.64 | 2.10 | 5.84 | 2.45 |
| CaO | .06 | 1.56 | .00 | .02 | .00 | .00 | .00 |
| BaO | n.d. | .06 | .32 | .03 | .10 | .08 | .28 |
| F | n.d. | .04 | .24 | .30 | .49 | .12 | .20 |
| -O=F | n.d. | -.02 | -.10 | -.13 | -.21 | -.05 | -.08 |
| Total | 94.31 | 95.25 | 95.65 | 97.25 | 95.98 | 93.89 | 94.42 |
| | | | | Cations based on 11 oxygens | | | |
| Fe | .03 | .03 | .17 | .06 | .09 | .04 | .08 |
| Mg | .01 | .00 | .09 | .00 | .04 | .00 | .02 |
| Mn | .00 | .00 | .00 | .00 | .00 | .00 | .00 |
| Ti | .01 | .00 | .01 | .00 | .01 | .00 | .01 |
| Al | 3.03 | 3.14 | 2.79 | 2.99 | 2.86 | 3.10 | 2.93 |
| Si | 2.97 | 2.85 | 3.01 | 2.99 | 3.04 | 2.94 | 3.00 |
| K | .08 | .10 | .81 | .00 | .67 | .12 | .62 |
| Na | .86 | .78 | .21 | .93 | .27 | .74 | .32 |
| Ca | .00 | .11 | .00 | .00 | .00 | .00 | .00 |
| Ba | n.d. | .00 | .01 | .00 | .00 | .00 | .01 |
| $(\frac{F}{F+OH})^1$ | n.d. | .01 | .05 | .06 | .09 | .02 | .04 |
| $\frac{K}{K+Na}$ | .09 | .11 | .79 | 1.00 | .71 | .14 | .66 |

n.d. = not determined
P = paragonite
M = muscovite
[1] calculated assuming (F + OH) = 2
[2] inclusion in garnet

Table 2. Mineral assemblages and selected compositions (from microprobe analyses) in paragonite-bearing rocks

| | Pa | Mu | Sill | Q | Gar | Ctd | St | Chl$^2$ | Gahn | Apat | Ilm | Hem | Rutile |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 77-163 | x | | | x | .97 | .95 | .90 | .88 | | | .95 | | |
| 77-244c | x | x | x | x | | | | | x | x | | .08 | x |
| 78-86b$^1$ | x | | | | .95 | | | | | | | | x |
| 81-42 | x | x | x | x | .94 | | .95 | | | | 1.00 | | x |

[1] inclusion in garnet

[2] retrograde

Numbers represent Fe/(Fe + Mg) for silicates, or $FeTiO_3/(FeTiO_3 + Fe_2O_3)$ for oxides

Al, a corresponding deficiency in Si and full occupancy of the alkali site, suggesting that Ca enters the structure through substitution toward margarite, $CaAl_4Si_2O_{10}(OH)_2$. Other samples have negligible amounts of calcium. Minor amounts of Fe and Mg occur in several samples, presumably occupying octahedral sites. They may enter the structure through the celadonite substitution ($Al^{VI} + Al^{IV} = (Fe, Mg)^{VI} + Si^{IV}$). Nearly all the paragonites have slight deficiencies in alkalis (*cf.* Henley, 1970; Hock, 1974; Baltatzis and Wood, 1977; Rumble, 1978) which may suggest the presence of hydronium ($H_3O^+$) in their alkali sites (Brown and Norrish, 1952). However, alternative interpretations such as substitution of $(Fe,Mg)^{VI}$ into normally unfilled octahedral sites, coupled with two vacancies on alkali sites, are also consistent with analytical data. Additional substituents include trace amounts of fluorine and barium. No paragonite had detectable amounts of chlorine.

Muscovite shows similarly large variations from its end-member composition of $KAlSi_3O_{10}(OH)_2$. The most abundant substituent is Na for K with up to 34 mole% of the potassium replaced by sodium (Table 1). Up to 3.9 wt.% (FeO+MgO) replaces $Al_2O_3$, presumably through substitution toward celadonite but possibly involving some $Fe_2O_3$ as well. The most celadonitic muscovite (77-244c) coexists with hematite, suggesting the possibility of $Fe^{3+}$ in the muscovite structure. Like paragonite, muscovite shows slight deficiencies in alkalis relative to its ideal chemistry. Trace amounts of Ti, Ba and F can be found in several samples, but none contain detectable concentrations of chlorine.

Other minerals found in these rocks are summarized in Table 2. Garnet forms solid solution between almandine and pyrope, with negligible amounts of Mn or Ca. Chloritoid, staurolite and chlorite consist of nearly pure Fe–Mg solutions, with less than 0.3 wt.% ZnO in staurolite of specimen 81-42 the only significant impurity. Gahnite has composition $Fe_{.08}Mg_{.10}Mn_{.02}Zn_{.80}Al_2O_4$ in the one sample in which it occurs. Sillimanite, quartz and rutile have compositions $Al_2SiO_5$, $SiO_2$ and $TiO_2$, respectively.

## Petrology

Two paragonite-bearing samples (77-163, 78-86b) come from rare garnet-rich veins that are exposed in cliffs of Ortega Quartzite, 1 km northwest of the summit of East Pecos Baldy in the south-central Truchas Range (Fig. 1). The veins cut across well-developed festoon crossbedding in the quartzite. Veins contain garnet, chloritoid, staurolite, quartz, white mica, apatite, tourmaline, large (0.1 mm), frosted, pitted and rounded zircon crystals, rutile and traces of retrograde chlorite. Veins consist of 90–95 modal percent euhedral garnet at their centers, grading into sieve-textured intergrowths of garnet, staurolite, chloritoid and quartz near their margins, then gradually grading into garnet-free Ortega Quartzite. Veins are up to 15 cm thick, irregular in shape, commonly anastomosing within the quartzite, and generally have a near-vertical orientation. Dozens of these veins occur at this locality. The origin of these veins is not well understood. A tentative interpretation is that they represent the metamorphosed equivalents of detrital clays and heavy minerals strapped in vertical fissures during or slightly after accumulation of the Ortega Quartzite. Such an interpretation is consistent with preliminary U–Pb studies of the zircons. The zircons yield an array of data points which indicate an age well in excess of 1700 m.y., the approximate depositional age of the quartzite (S. Bowring, pers. comm., 1983).

Regardless of their origin, the veins seem to have equilibrated at peak metamorphic conditions. Their mineral assemblages are compatible with those of the surrounding metamorphic rocks, and rare biotite-bearing veins give garnet-biotite temperature of 510°C using the geothermometer of Ferry and Spear (1978; *cf.* Grambling, 1983).

Paragonite does not seem to be in equilibrium with quartz in one of these samples (78-86b). Paragonite occurs only as inclusions in garnet and never touches quartz. The matrix of the sample contains muscovite, quartz and other minerals listed above, but no paragonite.

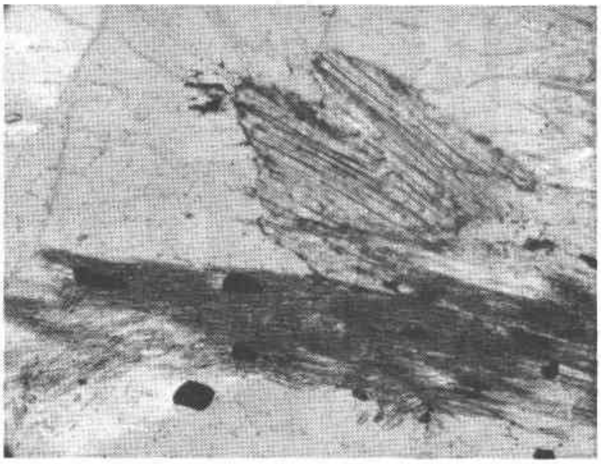

Fig. 4. Photomicrograph of coexisting paragonite, sillimanite and quartz in specimen 77-244c. Photograph is 1 mm long.

However, the other garnet-vein sample has paragonite in direct contact with quartz. No muscovite occurs in 77-163, but otherwise the sample appears similar to 78-86b. Minerals coexisting with paragonite and quartz are chloritoid, staurolite, garnet, zircon, rutile and trace amounts of retrograde chlorite which occur concentrated along garnet grain boundaries. The complete mineral assemblage is listed in Table 2. Paragonite forms crystals up to 3 mm in length, in places intergrown with chloritoid. Crystals are subhedral to euhedral, are never associated with chlorite, and do not appear to be retrograde in origin. Although sample 77-163 has no Al-silicates, it comes from the central part of the Truchas Range where kyanite, andalusite and sillimanite coexist (Fig. 2). Rocks with all three polymorphs of $Al_2SiO_5$ were collected 75 m north and 100 m south of this station. Evidence suggests that the rock has a stable association of paragonite + quartz and crystallized at conditions near the $Al_2SiO_5$ invariant point.

A third paragonite-bearing sample (77-244c) was collected from the eastern summit of Middle Truchas Peak (Fig. 1). The sample represents a slightly schistose, sillimanitic bed in Ortega Quartzite and contains muscovite, paragonite, quartz, sillimanite, hematite, rutile and scattered grains of bluish, Zn-rich spinel (Table 2). Sillimanite needles and hematite plates define the plane of $S_1$ foliation. Nearly all micas are oriented parallel to $S_1$. The two micas occur as separate crystals up to 1.5 mm in length, never as intergrowths, yet the two micas are in direct contact in several places and invariably lie within a few millimeters of each other. Both micas occur in direct contact with sillimanite and quartz, and intergrowths of paragonite, sillimanite and quartz are common (Fig. 4). No retrograde minerals can be found in the specimen and the rock seems totally unweathered. Textural observations suggest that the rock contains a stable assemblage of paragonite, quartz and sillimanite.

Specimen 77-244c was collected 4 km north of the regional kyanite-sillimanite isograd (Fig. 2). Its metamorphic temperature must have been somewhat higher than that of the $Al_2SiO_5$ triple point. Quartzites near Middle Truchas Peak contain coexisting chloritoid and sillimanite (Grambling, 1981, 1983), restricting temperature to a maximum of 560°C according to experimental work of Ganguly (1977). A reasonable estimate of metamorphic conditions would be 530–550°C, 4–4.5 kbar based on these constraints and the $Al_2SiO_5$ triple point of Holdaway (1971).

The fourth paragonite-bearing sample (81-42) comes from sillimanite-zone rocks in the Rio Mora region, 3 km west of the village of Gascon, New Mexico. This sample was collected from a 1 cm thick aluminous bed in Ortega Quartzite from outcrops at 3110 m elevation (Figs. 1, 2). Its assemblage consists of sillimanite–garnet-staurolite–paragonite–muscovite–quartz with minor ilmenite and rutile. Mineral compositions are listed in Table 2. Sillimanite includes coarse and fibrolitic crystals up to 3 mm in length, typically aligned parallel to $S_1$. Muscovite and paragonite form plates up to 4 mm across, with the two micas showing epitaxial intergrowths parallel to (001) (Figure 3). Most micas are oriented parallel to $S_1$. Paragonite and sillimanite commonly occur in direct contact, and quartz is ubiquitous. Again, textural observations indicate stable coexistence of paragonite, quartz and sillimanite.

The kyanite–sillimanite isograd forms a horizontal, nearly planar surface at Rio Mora. Sample 81-42 was collected from sillimanite-rich outcrops 300 m below the isograd, along the face of an 850 m-high cliff that bounds the eastern edge of the Rio Mora region. Grambling (1982) has estimated metamorphic conditions as 4.7 kbar, 550°C at the isograd. Such a temperature is consistent with the general absence of chloritoid from sillimanitic quartzite at Rio Mora (*cf.* Ganguly, 1977).

Although the three samples of paragonite–quartz come from widely separated areas, none has any textural features suggestive of disequilibrium. All have intergrowths of paragonite and quartz, and sillimanite forms an integral part of the mineral assemblage in two of them. The geometry of mapped $Al_2SiO_5$ isograds across the region (Grambling, 1981, 1982) requires that sillimanite was a stable phase in all paragonite localities. The dominant alignment of paragonite and sillimanite parallel to $S_1$ in two samples (77-244c, 81-42) provides textural evidence that the paragonite was present during $F_1$ deformation, so the paragonite cannot be a retrograde phase. Paragonite shows random orientation in one sample (77-163) and rare grains cut across $S_2$ in other samples (77-244c, 81-42), suggesting that paragonite partially recrystallized after $F_2$ folding. Therefore the paragonite did not behave as a non-reactive, relict phase. Several of the samples (77-163, 77-244c) have paragonite in crystals

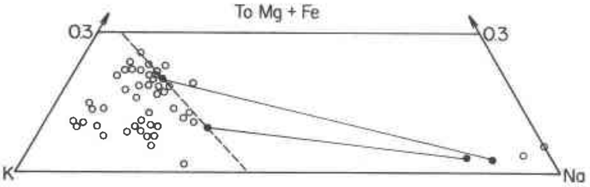

Fig. 5. Distribution of K, Na, and (Mg+Fe) in muscovite and paragonite. Data come from all analyzed muscovite–paragonite rocks which represent a temperature interval of about 520–550°C. Tie lines join the two muscovite–paragonite pairs, and the dashed line connects the two muscovites of these pairs. This dashed line approximately separates paragonite-free rocks from rocks with muscovite + paragonite. Its slope suggests that the solvus between muscovite and paragonite may broaden as (Mg+Fe) are added to muscovite.

totally separate from muscovite, suggesting that the paragonite did not form by exsolution during cooling. No paragonite-bearing rock contains plagioclase, an expected breakdown product of paragonite + quartz. It appears that the assemblage paragonite–quartz was stable under prevailing sillimanite-grade metamorphic conditions in these three localities.

### Solvus relationships

Sodic muscovite and potassic paragonite coexist in samples 77-244c and 81-42. Low-potassium paragonite (samples 77-163, 78-86b [as inclusions in garnet]) does not coexist with muscovite, and low-sodium muscovite (matrix, sample 78-86b) does not coexist with paragonite, consistent with the interpretation that mica compositions preserve evidence of equilibrium. Compositions of coexisting micas might be useful for geothermometry because coexistence of two white micas defines a solvus in the system $KAl_3Si_3O_{10}(OH)_2$–$NaAl_3Si_3O_{10}(OH)_2$.

Three different experimental and theoretical calibrations of the muscovite–paragonite solvus have been published recently (Eugster *et al.*, 1972; Thompson, 1974; Chatterjee and Froese, 1975). Samples reported in this study have compositions which do not plot on the solvus of Chatterjee and Froese (1975) for any reasonable pressures or temperatures. Samples do fall on the solvus of Eugster *et al.* (1972) or Thompson (1974). However, temperatures calculated from solvus relationships seem inaccurate. Temperature calculated from the muscovite limb in sample 77-244c (550°C) differs considerably from temperature calculated from the paragonite limb of the solvus in the same sample (620°C), using the Eugster *et al.* (1972) solvus. The other muscovite–paragonite sample (81-42) yields concordant but significantly higher temperatures of 675–680°C. Extrapolation of Thompson's (1974) solvus yields similar results. These temperatures were calculated at 2.07 kbar, so pressure corrections suggested by Thompson (1974) or Guidotti and Sassi (1976) would raise all temperatures by 30–50°C. Except for the 550°C temperature calculated from muscovite in 77-244c, all calculated temperatures are too high to be consistent with paragonite + quartz stability (Chatterjee, 1972), especially since K does not enlarge the stability of paragonite + quartz by more than a few degrees (Chatterjee and Froese, 1975). All but the 550°C temperature are also inconsistent with other observed mineral assemblages.

Previous studies of muscovite–paragonite pairs in natural samples have met with similar difficulties (Guidotti and Sassi, 1976; Baltatzis and Wood, 1977; Katagas and Baltatzis, 1980). Temperatures calculated from the muscovite limb of the solvus tend to disagree with temperatures calculated from the paragonite limb, and both temperatures may disagree with values calculated from other phase equilibria. Possible reasons for these discordant temperatures include deviations of natural samples from experimental mineral compositions, potential effects of hydronium ($H_3O^+$), poorly understood effects of variable pressure on the solvus (*cf.* Guidotti andd Sassi, 1976 with Chatterjee and Froese, 1975), possibly-erroneous experimental data and thermodynamic extrapolations from those data (Essene, 1982), and partial re-equilibration of muscovite–paragonite pairs during cooling.

The scatter of calculated solvus temperatures in the New Mexico rocks may be related to the considerable content of impurities in the analyzed micas, especially Fe and Mg in muscovite. Figure 5, a phase diagram separating the effects of K, Na and (Fe+Mg), illustrates this suggestion. The diagram includes data from all 56 analyzed specimens. It assumes that all micas crystallized at similar temperatures, thus neglecting the 530–560°C temperature range encompassed by the samples. It further assumes that $Fe^{3+}$ is not a major substituent and that other substituents such as Ca or Ti have negligible effect on the solvus. With only two exceptions the white micas separate into three distinct fields: muscovite only, muscovite + paragonite, and paragonite only. Although both muscovite-paragonite rocks crystallized at similar *P–T* conditions, tie lines between muscovite and paragonite seem to show considerable dependence on (Fe+Mg) in muscovite. The ratio K/(K+Na) of muscovite increases sharply and K/(K+Na) of paragonite may decrease slightly as (Fe+Mg) increases. The scanty data base is supported by the general absence of muscovite analyses from the region to the right of the dashed line in Figure 5. The data suggest that addition of a celadonite component to muscovite may effectively "broaden" the solvus between muscovite and paragonite. Such a conclusion is consistent with observations of Hock (1974) and Katagas and Baltatzis (1980) in other areas. The same phenomenon could also explain unusual observations made by Mohr and Newton (1983), who observed an abrupt narrowing of the muscovite–paragonite solvus across their kyanite isograd in North Carolina. The narrowing of the solvus corresponds to an abrupt decrease in (Fe+Mg) in muscovite, the same relationship observed in the present study.

Considering the meager data base presented here, it remains possible that some substituent besides (Fe+Mg) causes the observed broadening of the muscovite–paragonite solvus. However, Thompson *et al.* (1977) suggest that Ca has a negligible effect on K/(K+Na) in muscovite–paragonite pairs, and few other substitutions occur in the white micas from New Mexico. Further analytical work is underway to test these relationships.

In any case it appears that the anomalously high temperatures calculated from the muscovite–paragonite solvus are not caused by (Fe+Mg). Extrapolation of data in Figure 5 to a celadonite-free composition suggests that Fe, Mg-free muscovite coexisting with sodic mica should have about 42 mole% Na, which would give a solvus temperature of about 725°C (Eugster *et al.*, 1972), considerably too high for these rocks. The reason for the anomalously high temperature is unknown. Results of this and other studies (Baltatzis and Wood, 1977; Katagas and Baltatzis, 1980; Essene, 1982) suggest that muscovite–paragonite solvus geothermometry should not be used until problems of calibration and mineral composition are resolved.

## Conclusions

Paragonite and quartz coexist in apparent equilibrium in Precambrian metamorphic rocks of northern New Mexico. They occur in an area containing kyanite–andalusite–sillimanite and in two areas where sillimanite is the only polymorph of $Al_2SiO_5$ present. These are the first reported occurrences of assemblages with paragonite and quartz in a sillimanite-grade area, although Holdaway (1978) documented the occurrence of muscovite–paragonite–quartz in kyanite–andalusite rocks from the Picuris Range.

Because no samples contain the univariant assemblage paragonite–quartz–sillimanite–albite, it is difficult to determine an accurate temperature of metamorphism from the paragonite itself. However, coexistence of paragonite + quartz requires temperature less than 600°C at $P$ less than 5 kbar (Chatterjee, 1972). This maximum stability temperature is not significantly affected by substitution of K for Na, and solid solution of paragonite toward margarite or celadonite should reduce the thermal stability of paragonite + quartz (Chatterjee and Froese, 1975).

Chatterjee and Froese predicted that significant $Fe^{3+}$ might stabilize paragonite + quartz to temperatures above 600°C. Hematite coexists with paragonite in one specimen (77-244c), but the total iron content of that paragonite is only 0.6 wt.%. Other paragonites occur in more reduced assemblages and contain similar amounts of iron, so it is not clear that the Fe in paragonite is oxidized. Iron contents near 0.6 wt.% (as FeO) are common in natural paragonites (Henley, 1970; Baltatzis and Wood, 1977; Hoffer, 1978; Katagas and Baltatzis, 1980), and such negligible concentrations should not have a significant effect on experimentally determined stability curves, assuming Henry's Law behavior.

Temperatures below 600°C are supported by garnet–biotite geothermometry and by silicate phase relations (Holdaway, 1978; Grambling, 1981, 1982, 1983 and in prep.). The garnet–biotite geothermometer of Ferry and Spear (1978) yields temperatures of 500±25°C from Pecos Baldy, slightly south of paragonite localities 77-163 and 78-86b; 510±25°C from garnet veins at locality 77-163; and 550±25°C from Rio Mora. A small temperature difference between the southern Truchas Range and Rio Mora is consistent with differences in mineral assemblages between the areas. Even if the garnet–biotite geothermometer is incorrectly calibrated, the temperature differences should be real. Therefore, it seems certain that the paragonite + quartz rocks in the southern Truchas Range crystallized at least 50°C below the maximum stability of that assemblage, *i.e.*, below 550°C. Because surrounding rocks have sillimanite, the $Al_2SiO_5$ triple point must also lie below 550°C.

All mineral assemblages that can be considered diagnostic of the $Al_2SiO_5$ invariant point of Holdaway (1971) have now been found. Holdaway (1978) predicted that paragonite–quartz–sillimanite assemblages might be found in the Precambrian rocks of northern New Mexico, and this report substantiates his prediction. Other assemblages diagnostic of the Holdaway (1971) invariant point include chloritoid–sillimanite–quartz and kyanite–cordierite, and both occur in the area (Holdaway, 1978; Grambling, 1981).

It is likely that paragonite–quartz assemblages will be found in other areas where kyanite, andalusite and sillimanite coexist. The apparent scarcity of rocks with paragonite–quartz–sillimanite is in part an artifact of analytical techniques. It is expensive and time-consuming to perform microprobe and X-ray analyses on micaceous rocks, yet paragonite and muscovite have virtually identical optical and physical properties so are difficult to distinguish otherwise. The scarcity of this assemblage in schists may also result in part from a combination of $X_{H_2O} < 1$ and Ca-bearing bulk compositions.

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