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Petrological evidence for the development of refolded folds during a single deformational event

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Abstract—The Pelona Schist, which forms the lower plate of the Vincent thrust in the San Gabriel Mountains of southern California, has undergone a complex history of folding. The youngest folds in the schist (style 2 folds) range in shape from open to tight and fold both compositional layering and schistosity. These are superposed upon isoclinal folds with axial-plane schistosity (style 1 folds) that, in turn, overprint older isoclinal folds (also called style 1 folds). Samples from the hinges of style 2 folds contain two generations of muscovite. Muscovites of the older generation are parallel to the folded (style 1) schistosity. The newer muscovites recrystallized during and/or after style 2 folding. Microprobe analysis indicates that the two generations of muscovite are very similar in composition, although the new muscovites tend to have slightly higher paragonite and celadonite contents than the old muscovites. From the gross similarity of the two groups of muscovite, it is concluded that the style 1 and style 2 folds were produced during a single progressive deformation. The slightly higher paragonite and celadonite contents of the new muscovites are thought to indicate that both pressure and temperature were increasing during the deformation. This is consistent with the deformation being due to underthrusting of the Pelona Schist beneath the upper plate of the Vincent thrust.

INTRODUCTION

REFOLDED folds are widespread in orogenic belts. In the past, refolded folds have often been considered as evidence of discrete pulses of deformation. More recently, many workers have suggested that multiple generations of folds, particularly in thrust zones, can be the product of a single, progressive deformation (Bryant & Reed 1969, Williams & Zwart 1977, Wood 1978, Talbot 1979, Mattauer *et al.* 1981). The purpose of this study is to consider the origin of refolded folds in the Pelona Schist, which forms the lower plate of the Vincent thrust system of southern California and southwestern Arizona.

The Pelona Schist is composed predominantly of metagraywacke and mafic schist. It is overlain along the Vincent thrust by Pre-Cambrian, Palaeozoic, and Mesozoic gneisses and felsic plutonic rocks. The lower 1 km of the upper plate exhibits mylonitic textures. The schist was metamorphosed at relatively high pressures (6–7.5 kbar, Graham & England 1976) and is believed to represent a fossil subduction complex of late Cretaceous–early Tertiary age (Burchfiel & Davis 1981, Crowell 1981, Dickinson 1981; but see Frost & Martin (1983) for a different interpretation of the age of the Vincent thrust and Pelona Schist).

The structure of the Pelona Schist is complex; open to tight folds (style 2 folds) are superposed upon isoclinal folds (style 1 folds) that, in turn, refold older isoclinal folds (also called style 1 folds) (Jacobson 1980, 1983a, c). Early and late isoclinal folds are identical in both style and orientation, and can be distinguished only where overprinting is evident. It is for this reason that they both are referred to as style 1 folds. Interpretation of the folds is controversial, particularly regarding whether or not the style 2 folds formed during movement of the Vincent thrust and can be used to infer direction of transport (Haxel & Dillon 1978, Burchfiel & Davis 1981, Crowell 1981, Ehlig 1981). The goal of this paper is to shed some light on the controversy by studying the petrology of muscovites that grew during the different folding events.

Analysis of multiple generations of muscovite has proved helpful in a number of metamorphic terranes (Boulter & Råheim 1974, Chopin & Maluski 1980, Liewig *et al.* 1981, White & Johnston 1981, Saliot & Velde 1982). The usefulness of muscovite results from the fact that it undergoes solid-solution with the sodium mica paragonite ($NaAl_2AlSi_3O_{10}(OH)_2$) and the tetrasilicic mica celadonite ($K(Mg,Fe)AlSi_4O_{10}(OH)_2$). Paragonite content of muscovite has generally been found to increase with temperature, at least in rocks of low to medium metamorphic grade (Lambert 1959, Cipriani *et al.* 1971, Guidotti & Sassi 1976). Celadonite content tends to decrease with increasing temperature and increase with increasing pressure (Ernst 1963, Velde 1965, Sassi & Scolari 1974, Guidotti & Sassi 1976). Many authors have noted that the composition of muscovite varies not only with pressure and temperature but also significantly with mineral assemblage and therefore bulk-rock composition (Ernst 1963, Brown 1968, Guidotti & Sassi 1976). As shown below, these latter two controls are of some significance to the Pelona Schist.

This paper is an extension of an earlier study I conducted on muscovite in the Pelona Schist (Jacobson 1983b). The previous work emphasized the variation of muscovite composition with respect to distance from the Vincent thrust. A limited amount of data suggested that different textural groups of muscovite in the same sample might differ slightly in composition. The present study was conducted to test that possibility. The presence of textural groups of different composition is important because it enables one to infer the pressure–

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Table 1. Mineral percentages in the analyzed samples. Concentrations determined by counting 1000 points. 'X' indicates a mineral that, due to low abundance, did not fall directly under the cross-hairs during the point-counting. Qtz, quartz; mu, muscovite; ab, albite; ep, epidote; sph, sphene; chl, chlorite; act, actinolite; stp, stilpnomelane; cal, calcite; bio, biotite; gar, garnet. 'Other' includes accessories (tourmaline, zircon and apatite) and secondary vein minerals (quartz, calcite, chlorite, and potash feldspar)

Sample	qtz	mu	ab	ep	sph	chl	act	stp	cal	bio	gar	other
394	36.6	21.8	22.9	6.2	1.7	6.2	X		4.0		X	0.6
394a	35.2	20.6	24.2	6.6	2.2	8.9	0.4		1.3		0.1	0.5
462b	22.6	32.5	31.2	3.0	1.3	8.0				1.0		0.4
421	43.5	41.6	5.2	8.4	0.7	X	0.1	X		X	X	0.5
421b	29.1	22.9	37.7	3.2	1.0	5.4	0.2		X		0.3	0.2
421h	20.9	40.3	22.3	3.7	2.4	6.9	2.6					0.9
421i	26.5	31.7	30.5	4.3	1.9	4.3		0.5				0.3
421m	35.3	31.7	21.2	3.6	1.6	3.2	2.6	0.3	X			0.5
421n	18.6	50.3	13.5	6.3	2.4	1.5	1.3	5.5				0.6

temperature (P–T) history at individual localities. In particular, the present study provides information about pressure history that could not be obtained from the first study.

DESCRIPTION OF THE SAMPLES

The area of study is located in the San Gabriel Mountains of the Transverse Ranges of southern California (see location maps in Ehlig 1968, Haxel & Dillon 1978, Crowell 1981, Jacobson 1983a, b). In this region, a structural thickness of 4 km of Pelona Schist is exposed beneath the mylonites of the Vincent thrust zone. The base of the schist is not exposed. Style 1 folds (isoclinal folds and isoclinally-refolded folds) occur throughout the schist. These folds have an excellent axial-plane schistosity. Style 2 folds, which fold the style 1 schistosity, occur mostly within the 700 m of structural section directly subjacent to the thrust. None were found farther than 1000 m structurally below the thrust. The axes of the style 1 and style 2 folds are parallel.

Nine samples from style 2 fold hinges were collected for microprobe analysis. The samples were divided into two groups, A and B. Group A consists of three samples collected relatively close to the thrust (samples 394 and 394a from 50 m below the thrust and sample 462b from 175 m below the thrust). Group B consists of six samples from a single outcrop (locality 421) situated 1000 m below the thrust.

The sampled folds are relatively tight. Most have interlimb angles of 45° or less. The folds exhibit a crenulation caused by crumpling of the folded style 1 schistosity. A secondary cleavage parallel to the axial planes of the crenulations is moderately developed in two samples and weakly developed in several others. In a few samples, the crenulations are accompanied by a weak, axial-plane metamorphic segregation defined by layers enriched alternately in muscovite and quartz.

In all samples, muscovite occurs in two forms: (1) as old grains that grew during isoclinal folding and that were crenulated during the style 2 folding and (2) as new, strain-free grains that formed by recrystallization of the crenulated muscovites.

The old muscovites occur in elongated clots defining open to tight crenulations. In some cases, individual grains within the clots curve smoothly around the crenulation hinges. In other instances, grains on opposite limbs of a crenulation bend towards the hinge, where they meet at a high-angle grain boundary. Such evidence of grain-boundary migration is best developed where the crenulations are tightest. The old muscovites commonly contain fine inclusions of graphite.

In contrast to the deformed, old muscovites are the relatively strain-free, new muscovites. These occur as single crystals cutting across crenulated muscovites, as aggregates of randomly oriented grains partly to nearly completely replacing clots of deformed muscovite, and as single crystals surrounded by quartz and/or albite. The new muscovites are optically strain-free. They are generally devoid of inclusions and typically have low aspect ratios (3:1 or less). Representative photomicrographs of both old and new muscovites are shown in Jacobson (1983b).

Many muscovites show textures intermediate between those of the ideal old and new grains. Such grains were avoided during the microprobe analysis.

The complete mineral assemblages of the nine samples are indicated in Table 1. Besides muscovite, all samples contain quartz, albite, epidote, chlorite and sphene. In addition, all except two contain actinolite. Stilpnomelane, calcite, biotite and garnet are less common. The garnet is intermediate in composition between almandine, grossular, and spessartine (Jacobson 1980, 1983b). Thus, all assemblages indicate metamorphism of greenschist facies.

Of the above minerals, quartz, chlorite, stilpnomelane, calcite, and biotite are relatively strain free. Of these, all but biotite are common in rocks that have not undergone style 2 folding (Jacobson 1980) and it is therefore assumed that their strain-free nature is due to recrystallization during style 2 folding rather than to neomineralization. In contrast, biotite is relatively uncommon in the Pelona Schist of this region and is present in only two of the samples analyzed here (Table 1). It is thus not clear whether the strain-free habit of biotite indicates recrystallization or first growth during style 2 folding. Albite and epidote apparently did not undergo substantial recrystallization at this time, as indicated by inclusions of quartz and graphite that define a relict style 1 internal schistosity. The albite and epidote, however, do exhibit small overgrowths which may have formed during style 2 folding. Actinolite is zoned and may have grown throughout much of the complex deformational history. The timing of garnet growth is not clear.

MICROPROBE DATA

Microprobe analyses were performed on both old and new muscovites in all nine samples. The work was conducted at the University of Wisconsin, Madison on an ARL-SEMQ microprobe with Tracor Northern automation. Wavelength-dispersive spectrometers were used for all elements. Accelerating voltage and sample current were 15 kV and 16 nA, respectively. The beam was defocused to a diameter of approximately 7 μm. Most elements (Na, K, Si, Al, Fe and Mg) were analyzed for 40 seconds, or until sufficient data were obtained to reduce the X-ray counting error (1.0 standard deviation) to one-half per cent relative. For the minor elements (Ca, Mn, Ba and Ti), counts were collected for 5 or 40 seconds, depending on the peak-to-background ratio.

All spots analyzed were marked on photomicrographs of the probe sections. This made it possible to study the sections at a later date with a petrographic microscope to insure that no inclusions had been inadvertently analyzed.

To check for volatilization, trial runs were made in which muscovite was placed under the beam for five minutes and counts were collected for sodium and potassium. No decrease of count rate was noticed with time. During the actual analyses, counts for Na and K were collected within the first minute of beam time.

Standards were natural silicates and oxides. The data were corrected using the Bence–Albee procedure.

The results of the microprobe study are shown in Fig. 1. Data are plotted for nine of the ten elements determined (Ca is not present in significant quantities). Average values of old grains are indicated by closed circles, new grains by open circles. The error bars represent one standard deviation of the mean (i.e. the standard deviation of the sample population divided by the square root of the number of analyses). The number of analyses per sample is indicated in the caption of Fig. 1. Atomic proportions were calculated on an anhydrous basis of 11 oxygens per formula unit. Iron was assumed to be completely in the ferrous state.

The data show a number of distinctive features.

(1) For all samples, the new grains have a higher average Na content than the old grains. (It should be noted that all six samples of Group B were collected within several feet of each other. The order of the samples on the horizontal axis of Fig. 1 reflects nothing other than the random order in which the samples were collected. Thus, the fact that the Na values of the Group B samples define a regular, arch-like pattern is completely coincidental).

(2) Group A muscovites are higher in Na than Group B muscovites. This is true for both old and new muscovites and confirms previous observations of an increase of Na content near the Vincent thrust (Jacobson 1983b).

(3) In all samples, Na and K sum to less than 1.0. The minimum value is 0.90 for old muscovite in sample 421h and the maximum is 0.97 for new muscovite in sample 421. Since H_2O was not determined, it is not known whether the deficiency of Na + K is due to vacancies in the interlayer-cation site or to the presence of H_3O^+. Part of the deficiency can be accounted for by the presence of Ba and possibly by substitutions involving Ti. As noted previously, there does not appear to have been significant volatilization of Na and K during analysis. Deficiencies of interlayer-cations in muscovite are not rare (Ernst 1963, Brown 1967, Butler 1967, Black 1975, Katagas & Baltatzis 1980).

(4) In seven of the samples, the increase of Na in the new muscovites is accompanied by an increase of K. This indicates that the new muscovites must be lower in either interlayer vacancies or H_3O^+.

(5) Silicon in all cases exceeds the 3 atoms per formula unit of an ideal muscovite due to celadonite substitution. Group A and Group B each contain one sample that is notably high in Si compared to the rest of the samples of the group (sample 462b in Group A and sample 421 in Group B). Samples 462b and 421 are the only two containing biotite (although in neither sample is it abundant). This is consistent with previous observations that muscovite in biotite-bearing assemblages tends to have a high celadonite content (e.g. Ernst 1963). Particularly when the role of biotite is taken into account, Group A samples are seen to be lower in celadonite than Group B samples. This confirms previous observations (Jacobson 1983b).

(6) Old and new muscovites show no systematic difference in Si content. In three samples (394a, 421h, 421n), Si contents of old and new grains differ by less than one standard deviation of the mean. For the remaining six samples, the differences are only slightly greater than one standard deviation. In four of the six, the new muscovites are lower in Si than the old ones. In the remaining two samples, the opposite is the case.

(7) As expected by the nature of celadonite substitution, the relations for Al are essentially opposite to those of Si. That is, samples high in Si (e.g. 421) are low in Al and vice versa (e.g. 394 and 394a). In addition, in all samples for which the old grains are higher in Si than the new grains, the old grains are lower in Al (and vice versa).

(8) For eight of the nine samples, the sum of Mg and Fe is lower in the new muscovites than the old ones. This consistency is surprising, because, as just noted, Si shows no systematic difference between old and new grains. Furthermore, Mg, Fe, and Si in celadonitic muscovite should obey the following relation:

$$\mathrm{Mg} + \mathrm{Fe} = \mathrm{Si} - 3. \qquad (1)$$

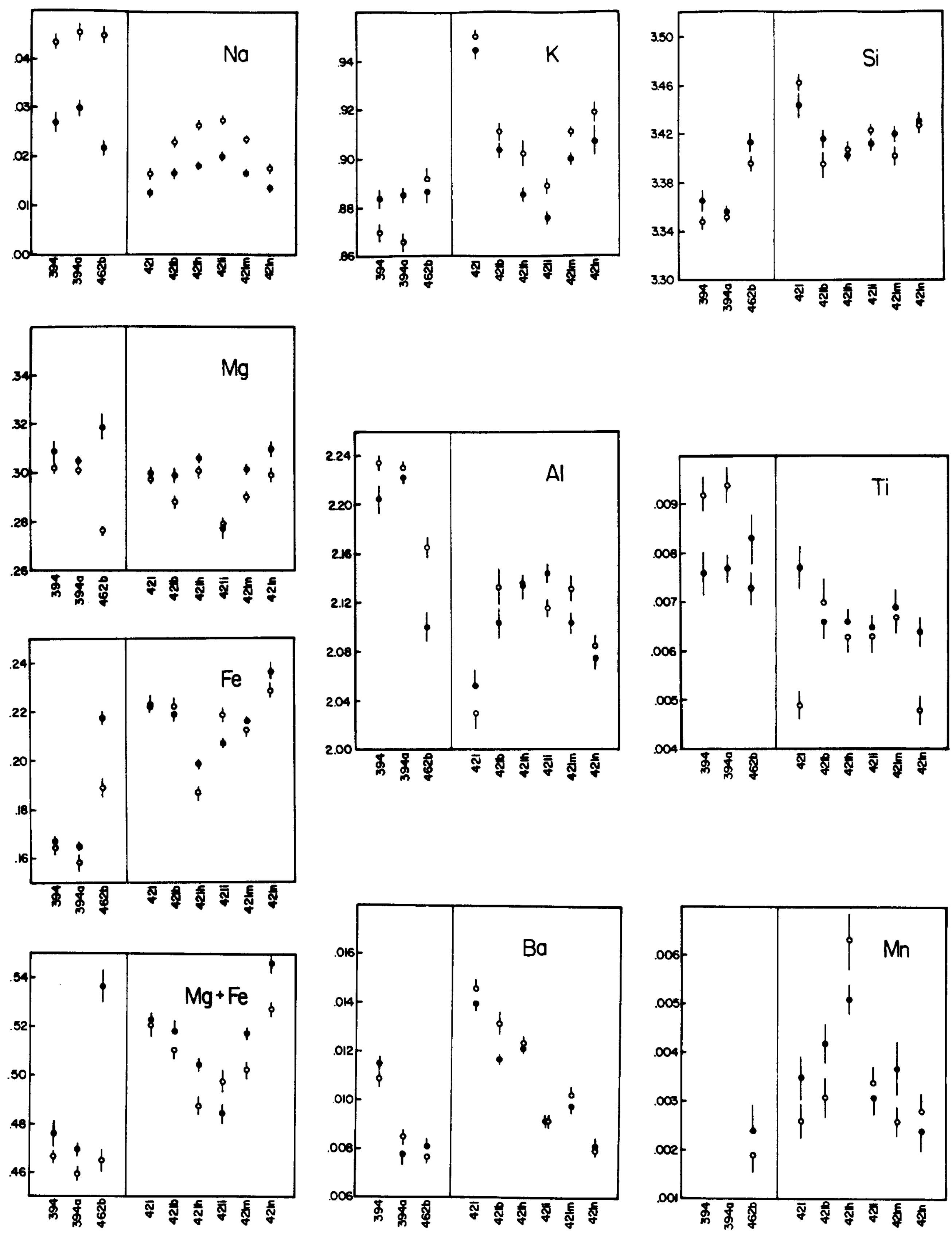

Fig. 1. Composition of old and new muscovites. Concentrations are in moles per 11 oxygens. All iron as FeO. Old grains indicated by closed circles, new grains by open circles. Group A samples lie to the left of the vertical line, Group B samples to the right. Error bars indicate one standard deviation of the mean. The number of analyses for each sample is as follows: 394(old-19, new-33), 394a(64,60), 462b(31,37), 421(36,61), 421b(34,30), 421h(51,29), 421i(38,38), 421m(35,59), 421n(30,53).

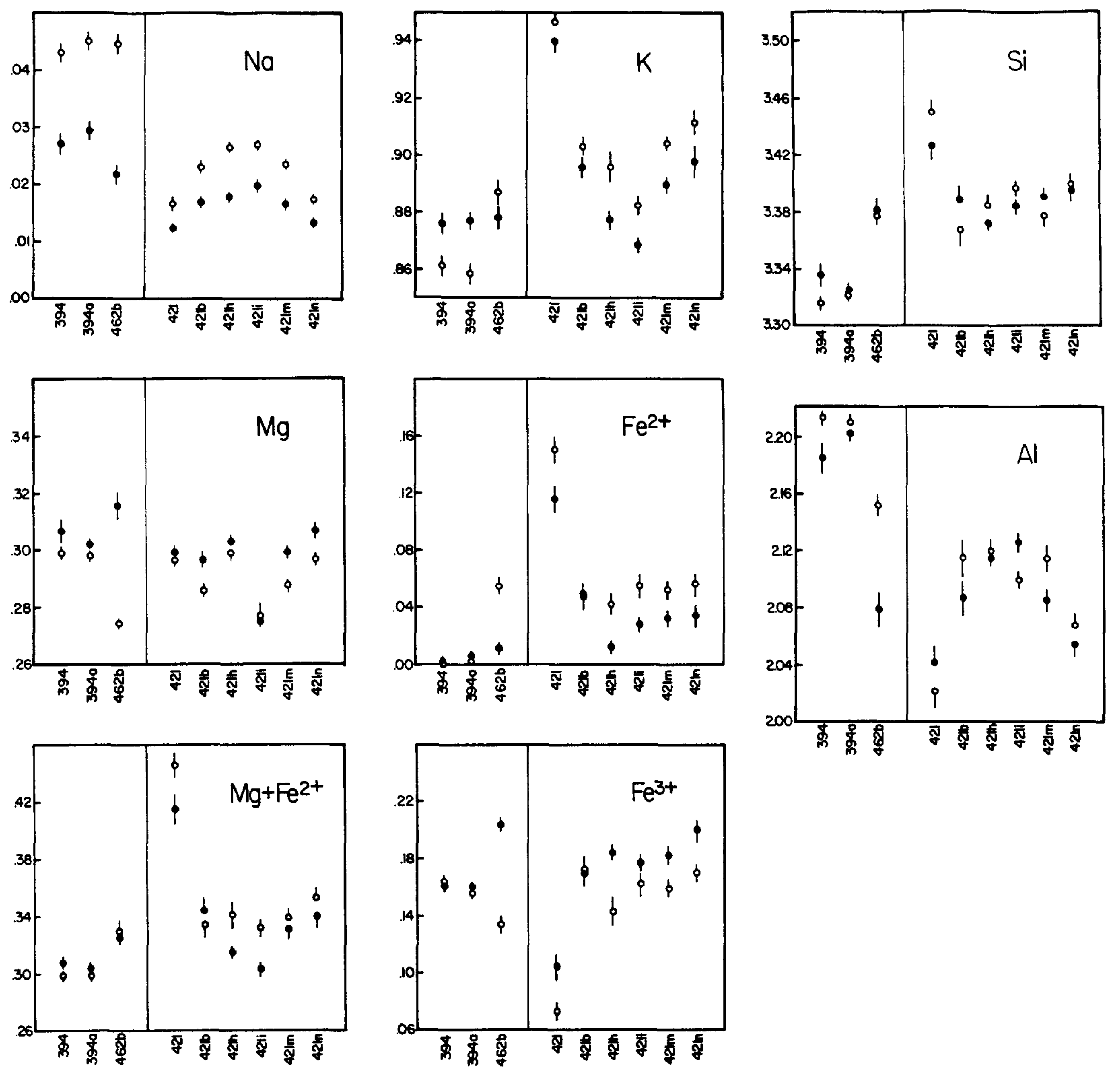

Fig. 2. Selected data from Fig. 1 corrected for ferric iron (see text). Error bars (one standard deviation of the mean) indicate the variability of the individual corrected analyses but provide no information about systematic errors involved in the recalculation procedure.

The plots of Si and Mg + Fe (Fig. 1), however, indicate that total Mg + Fe is substantially greater than that necessary to explain the excess of Si atoms above the three present in ideal muscovite.

A plausible explanation for the above inconsistencies is that they result from the assumption that Fe is present only in the ferrous state. For this reason, all analyses were recalculated to estimate the amount of ferric iron present. Each analysis was corrected by multiplying the mole proportions of all elements by the factor (always less than 1.0) that would force the tetrahedral and octahedral cations (Si, Al, Mg, Fe(total), Ti and Mn) to sum to 6. This caused the number of oxygens per formula unit to drop below 11. Ferric iron was then calculated to be that amount necessary to restore the number of oxygens to 11. A sum of 6 for the tetrahedral and octahedral cations implies the complete absence of solid solution with a trioctahedral mica. Such substitution is generally thought to be small but not absent (Foster 1956, Brown 1967, 1968, Katagas & Baltatzis 1980). Thus, the recalculation procedure may overestimate the amount of ferric iron present.

The recalculation procedure results in a uniform reduction of all cations in an analysis. The magnitude of reduction for each sample depends upon the uncorrected sum of octahedral and tetrahedral cations. The higher that sum, the greater the reduction and the greater the estimated content of ferric iron. Most samples have approximately the same sum of octahedral and tetrahedral cations and have undergone similar corrections. For this reason, the corrected plots for Na, K, Si, Al and Mg (Fig. 2) are very similar in overall pattern to the uncorrected plots (Fig. 1), except that all points are shifted downward relative to the vertical axis. Sample 421,

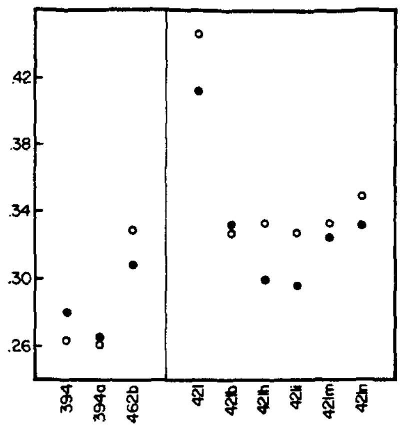

Fig. 3. Mole per cent celadonite (moles Si − 3.0 − moles pyrophyllite). The data are corrected for ferric iron. It is assumed that no H_3O^+ is present in the interlayer position.

which has the lowest estimated content of ferric iron, is shifted the least. The plots of Fe^{2+}, Fe^{3+}, and Mg + Fe^{2+} (Fig. 2) obviously differ from any plots in Fig. 1 because of the separation of ferrous and ferric iron.

Figure 2 shows that the new muscovites are, on average, lower in Fe^{3+} and higher in Fe^{2+} than the old muscovites. Figure 2 additionally indicates that, in all samples, the sum of Mg + Fe^{2+} is less than the excess of Si atoms above three. This is opposite to the situation noted for the uncorrected sum of Mg + Fe (Fig. 1). One possibility is that Fe^{3+} has been overestimated, leaving too little Fe^{2+}. A second possibility is that the high values of Si are due not only to celadonite substitution but also to pyrophyllite substitution ($\square Al_2Si_4O_{10}(OH)_2$). The presence of pyrophyllite substitution also would explain the previously noted minor deficiency of interlayer cations. Thus, a second correction was made to estimate the celadonite content of each sample by subtracting the interlayer cation deficiency from the excess of Si atoms above three. The results are shown in Fig. 3. This figure should be compared to the plot of Mg + Fe^{2+} in Fig. 2. If the two figures are overlaid, the sample points nearly coincide (particularly for the six samples of Group B). This is expected from equation 1 and is evidence that the various assumptions made in the recalculations are reasonable. Figure 3 shows that, in six samples, the new muscovites are higher in celadonite than the old ones. Three samples show a decrease of celadonite for the new muscovites, but in two of the cases the decrease is small. No error bars are shown in Fig. 3 because they would not reflect the probably large systematic errors involved in the several assumptions used to obtain the plotted values.

Particular note should be made of samples 421 and 462b. These are the two biotite-bearing assemblages. As stated previously, it is not clear from textural relations whether the biotite first grew during style 2 folding or whether it was present also during style 1 folding. If it first appeared during style 2 folding, then the muscovites that grew contemporaneously (i.e. the new muscovites) should be lower in celadonite than the old muscovites (Ernst 1963 fig. 2). In fact, both samples show an increase of celadonite content in the new muscovites. This implies either that biotite *was* present during style 1 folding or that the inferred values of celadonite content are incorrect.

DISCUSSION

The data of Figs. 1 to 3 show that Group A samples tend to be higher in sodium and lower in celadonite than Group B samples. These trends have been described elsewhere and are thought to indicate a slight increase of temperature with approach to the Vincent thrust (see Jacobson 1983b).

Most important for this study is a comparison of old and new muscovites within the same sample. Several elements do show minor differences in concentration between old and new muscovites (e.g. Na is consistently higher in the new muscovites). However, when the data are viewed as a whole, the most striking feature is that old and new muscovites within a given sample are actually quite similar. For example, the greatest difference in Si content is shown by sample 421b (Fig. 1). In this case, the old muscovites have an average silicon content of 3.416 ± 0.008 moles per formula unit, the new muscovites an average of 3.396 ± 0.010 moles (Fig. 1) for a difference of 0.02 moles. In the case of Mg + Fe, old and new muscovites differ by less than 0.02 moles in all cases except sample 462b. Similar relations are found for other elements. Particularly when considering the great range in composition shown by muscovites from various metamorphic terranes (Cipriani *et al.* 1971), it appears that the old and new muscovites must have crystallized under very similar conditions. Except by coincidence, this would seem to indicate that the two groups of muscovites grew within a short time of each other during the same metamorphic event. Since the old and new muscovites are associated with the style 1 and style 2 folds, respectively, it is concluded that the two sets of folds must be closely related in time.

It is assumed above that the compositional similarity of old and new muscovites is not due to chemical re-equilibration of the old muscovites at the time of formation of the new muscovites. That this is a reasonable assumption is indicated by the low grade of metamorphism (greenschist facies) and by other studies of greenschist and higher grade terranes in which old and new muscovites have been found to exhibit significant differences in composition (Boulter & Råheim 1974, Liewig *et al.* 1981, Saliot & Velde 1982).

One point that is not clear is whether the compositions of the old muscovites reflect the metamorphic conditions during formation of both early and late style 1 folds, or whether pervasive recrystallization during formation of the late style 1 folds caused evidence of an earlier, different metamorphism to be erased. Thus, the data presented above apply with certainty only to the relationship between the style 2 folds and the late isoclinal (style 1) folds. It is tempting to conclude that the early isoclinal folds are part of the same progressive deformational event, but this is not proven.

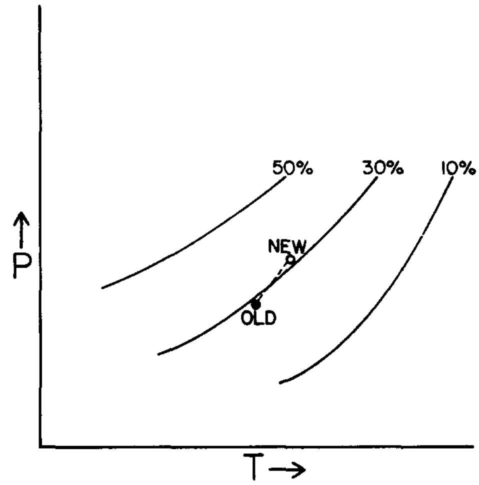

Fig. 4. Inferred pressure–temperature path for the Pelona Schist during style 2 folding. Isopleths of celadonite content are schematic and are based on Fig. 1 of Velde (1967).

Old and new muscovites are similar in composition but not identical. It is interesting to consider what the small differences mean. The most consistent difference is in Na content, which in all samples is higher in the new muscovites. It has often been noted for low-grade metamorphic rocks that Na content of muscovite tends to increase with increasing temperature (Cipriani *et al.* 1971, Guidotti & Sassi 1976). In many cases, this is due to increased miscibility with coexisting paragonite (Rosenfeld *et al.* 1958, Zen & Albee 1964). Increase of Na with temperature has also been noted in various assemblages not containing paragonite (Lambert 1959, Ernst *et al.* 1970, Cipriani *et al.* 1971, Guidotti & Sassi 1976), although none of these studies dealt with mineral assemblages precisely like those present in the Pelona Schist. Nevertheless, the fact that all nine samples, collected from several outcrops and exhibiting a variety of mineral assemblages, show the exact same trend appears at least suggestive of a slight rise in temperature during style 2 folding.

In contrast to Na, Si shows no consistent variation between old and new grains. When Si is corrected for pyrophyllite substitution (Fig. 3), new grains do show a slight tendency to be enriched in celadonite relative to the old grains. Studies of the effects of pressure and temperature on celadonite substitution are abundant and indicate a positive correlation with pressure and a negative correlation with temperature (Ernst 1963, Velde 1965, Mather 1970, Di Pierro 1973, Sassi & Scolari 1974, Guidotti & Sassi 1976). Although experimental calibration of this solid-solution series has been worked out only for the assemblage of muscovite + biotite + K feldspar + quartz (Velde 1965), systematic behavior of celadonite substitution in muscovite has been observed for many assemblages, including ones similar to those of the Pelona Schist (Sassi & Scolari 1974, Fettes *et al.* 1976, Briand 1980). The proposed temperature increase during style 2 folding should thus have favored a decrease of celadonite content. The fact that celadonite content in a majority of samples is higher in the new muscovites would seem to require a concurrent pressure increase to counterbalance the temperature effect. The proposed P–T path is shown in Fig. 4. The isopleths of celadonite content are based on the experimental work of Velde (1965, 1967). Strictly, this work applies only to the assemblages of muscovite + biotite + K-feldspar + quartz. However, it is suggested here on the basis of the empirical studies just cited that an approximately similar distribution of isopleths should apply also to the assemblages present in the Pelona Schist.

It must be re-emphasized that, because of the effects of mineral assemblage and because of possible systematic errors in the corrections for ferric iron and pyrophyllite substitutions, the inferred increases of both temperature and pressure are highly uncertain. Nevertheless, they are consistent with the metamorphism being due to increasing depth of burial. This is precisely what is expected if both style 1 and style 2 folds were formed during underthrusting of the Pelona Schist beneath the upper plate of the Vincent thrust.

CONCLUSIONS

The Pelona Schist of the San Gabriel Mountains has undergone a complex history of folding. The data presented here suggest that the various 'generations' of folds were produced in a single event related to underthrusting of the schist along the Vincent thrust. I envisage a continuous and repetitive process of folding, stretching of folds, and refolding related to the thrusting (cf. Bryant & Reed 1969, Williams & Zwart 1977, Wood 1978, Talbot 1979, Mattauer *et al.* 1981). The intensity and prolonged nature of the deformation can be explained by the fact that the Vincent thrust is a relict subduction zone in which oceanic rocks (graywackes and basalts) were transported a great lateral distance beneath continental basement rocks (possibly 200 km, Burchfiel & Davis 1981). The proposed model implies that both style 1 and style 2 structures may contain information about thrusting direction. However, some problems associated with previous attempts to use style 2 folds for this purpose are discussed elsewhere (Jacobson 1980, 1983a). Additional structural studies are presently being conducted to help determine thrusting direction, since that direction is important for regional geological reconstructions.

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