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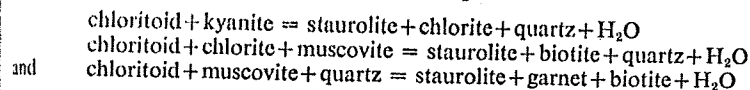
Three-dimensional isograds from the Lukmanier Pass, Switzerland, and their tectonic significance

J. S. FOX

Summary. Four mineralogical zones, corresponding to the following 'maximum' metapelitic assemblages

- zone 1: chloritoid-chlorite-kyanite
- zone 2: staurolite-chloritoid-chlorite-kyanite
- zone 3: biotite-staurolite-chloritoid-chlorite-kyanite
- zone 4: biotite-staurolite-garnet-chlorite-kyanite

have been mapped in the Ultrahelvetetic rocks of the Lukmanier area in Switzerland. Petrographic study suggests that, despite the ubiquity of 'disequilibrium', the zone boundaries accord with the simple univariant equilibria



Using structural contours, minimum isograd dips of 35° north have been obtained. The Lukmanier parageneses require the Lepontine phase of Alpine metamorphism to have had a geothermal gradient distinct from both Alpine blueschists and the Bergell contact metamorphites. The high minimum value for isograd and isotherm orientation here points to a convective heat source for the Lepontine event.

1. Introduction

Intensive study in almost every discipline of geology over the past decade has resulted in relatively detailed information on the complex plate interactive zone that is the Alpine orogen. Today the various Tethyan lithospheric plates operative, their kinematics and their relationships to global interactions elsewhere can be defined with some confidence (e.g. Laubscher, 1969, 1970, 1971; Smith, 1971; Dal Piaz, Hunziker & Martinotti, 1972; Dewey, Pitman, Ryan & Bonnin, 1973). The lack of precise data, and particularly data that can be related to orogenic paleo-therms has, however, prevented a more positive contribution to the Alpine story by metamorphic petrologists. Fundamental metamorphic problems remain essentially unresolved despite the excellent regional compilations available (E. Niggli, 1970; E. Wenk, 1970; Ernst, 1973; Zwart, 1974) and despite much debate. The dichotomous schools of thought on the genetic relationships between the three or more thermal events of Alpine orogenesis are nowhere more sharply divided than over the source of heat for the kyanite-sillimanite facies series, early Tertiary, Lepontine episode.

The obvious approach to the unanswered questions of Alpine metamorphism through detailed, three-dimensional mineralogical mapping which, despite the eminent suitability of the terrain, is only now commencing (Streckeisen & Wenk, 1974; P. Thompson, in the press). During the past four years I have studied medium-grade mineral zonation in the Lukmanier Pass area of South-Central Switzerland as a contribution to the required metamorphic coverage. The

Lukmanier area, located 75 km SE of Lucerne astride the cantons of Ticino and Grisons, embraces the thickest post-Hercynian section through medium grades of the Lepontine phase anywhere in the Swiss Alps and must therefore be considered critical to its study.

Sampling in the Lukmanier area during the summers of 1971–2 was restricted to pelitic rocks. Only the freshest possible metapelites and those showing the largest number of mineral phases were collected. Over 700 meta-pelitic samples examined represent a sampling interval of about one for every 20 m of sampling horizon. Detailed petrographic study was made of 500 specimens and microprobe analysis was undertaken on 9.

2. Geology

Although mineral zones and isograds have not been investigated until now, the Lukmanier area has attracted the interest of petrologists for over a century. Notable contributions were those of Holst Pellekaan (1913, unpubl. thesis, Zurich Univ.), Krige (1918) and Bosshard (1929). The classical geochemical reconnaissance of the Gotthard Massif cover of P. Niggli (1929) which included this area, has now been complemented by the modern geochemical and stratigraphic treatment of M. Frey (1969), J. D. Frey (1967), Chadwick (1968) and Sibbald (1971, unpubl. thesis, London Univ.) have elucidated the relationship between structure and metamorphism.

The attraction of the Lukmanier area lies not only in its spectacular mineralogy but in its relative geological simplicity. The Mesozoic, Ultrahelvetic marls, semipelites and pelites which underlie the pass seem to have been subject only to the Cenozoic Alpine orogeny. Although folded and thrust into east-plunging, north-dipping antiforms and synforms during four phases of Alpine deformation, the peak of metamorphic mineral growth was essentially post-kinematic; tectonism has thus only weakly disturbed the disposition of the metamorphic mineralogy. In contrast, both the para-autochthonous crystallines of the Gotthard massif, which bound the area of study to the N, and the schists and gneisses of the Pennide zone, which bound it to the S and E, include rocks deformed before the Alpine orogeny. Both preserve pre-Alpine paragenetic remnants (Steiger, 1962; Arnold 1970*b*; Hunziker, 1970) and in the case of the former, both Hercynian (Carboniferous-Permian) and Caledonian (Ordovician-Silurian) orogenic ages have been obtained (Arnold, 1970*a*).

The interesting meta-pelitic mineralogy which is the subject of this paper is found at the boundary between the Ultrahelvetic (lower Jurassic) and Quatern (upper Triassic) formations (Fig. 1). Although Ultrahelvetic pelitic and semipelitic compositions are found at other horizons in this area (the upper Quatern bears spectacular hornblende-staurolite and hornblende-chloritoid assemblages which are the subject of a paper in preparation), only the 10–15 m thick, graphitic, black, calc-free greywackes and schists of the basal Stgir provide a widely distributed, virtually isochemical sampling horizon. Bulk chemical analyses (Fox, 1974, table 4) show this horizon to be peraluminous, with alumina contents often exceeding 30% by weight.

The basal Stgir can be found locally over 1100 m of elevation and over 6 km of

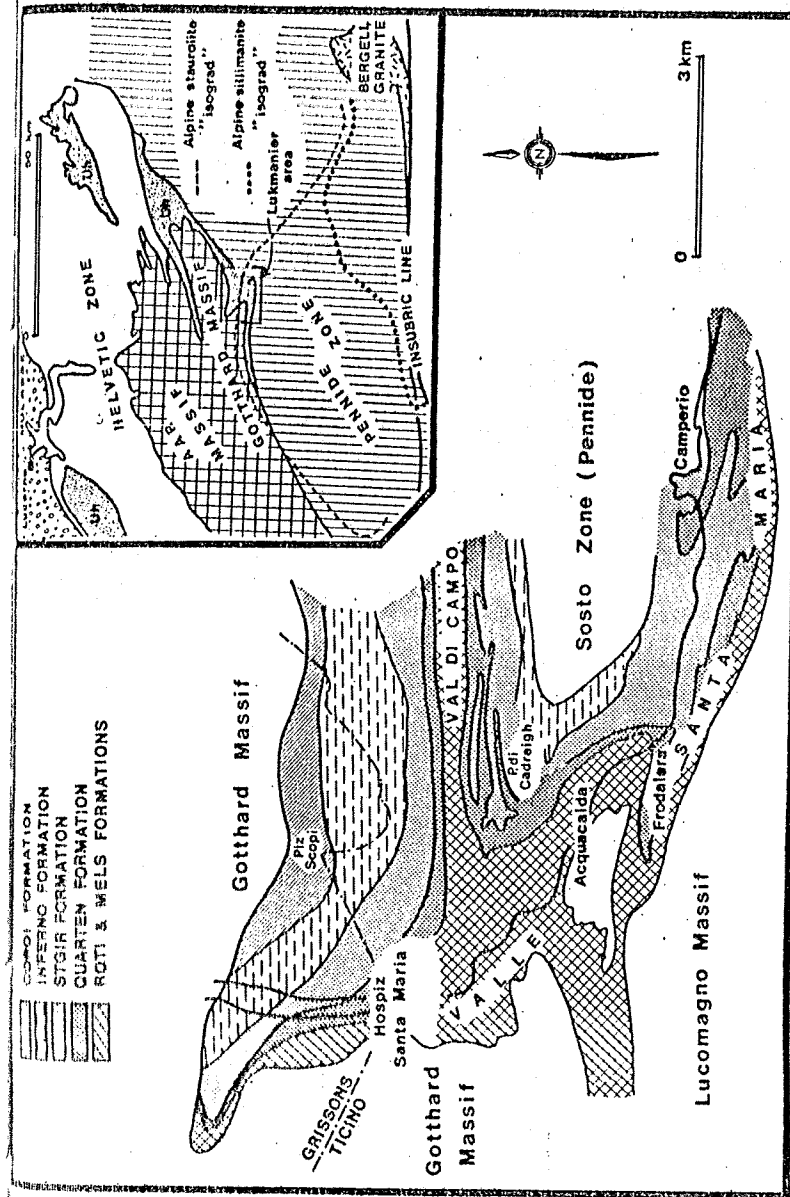


Figure 1. Geology of the Lukmanier Pass area, essentially after Chadwick (1968), and its relation to the Swiss-Alpine tectonic zones and to Niggli's (1970) Alpine isograds (uh = ultra helvetic).

N-S extension (equivalent to a minimum of 3.4 km of true 'thermal section', or about 65–70 °C – see Section 4). The presence of a 2–5 m thick, rusty weathering, massive ortho-quartzite directly underlying the pelitic sampling horizon and the abrupt transition upwards from the greens and buff's of the Quarten into the greys and blacks of the Stgir aid its field identification.

3. Mineralogical zonation

Detailed sampling of the basal Stgir has uncovered four metamorphic mineral zones in the area. The selection of samples was restricted to carbonate-free rocks in which powdery graphitic material was present in the matrix and which contained any combination of the minerals chlorite, chloritoid, biotite, staurolite, garnet and kyanite. The zones were defined so as to embrace rocks in which the maximum number of 'coexisting' phases among these six Fe-Mg-Al silicates represented a characteristic assemblage. The zone boundaries were drawn where the first of each of the four 'maximum' assemblages first appeared.

The defining samples in each zone were required to contain both quartz and white mica – most also contained zoned clinozoisite/epidote, tourmaline, zircon and apatite. Plagioclase was found to be a common phase in many of the higher grade specimens. Where such obvious bulk inhomogeneities as bedding laminae, coarse bedding or bedding-derived boudins were present in a thin section, the defining minerals were required to 'coexist' within a recognizable compositional realm.

Figure 2 shows the distribution of the 'maximum' assemblages, and the zone boundaries which they define, over the area. The method for extending boundaries from control points is discussed below. For comparison the approximate course of Niggli's (1970) staurolite 'isograd' is also illustrated.

3a. Zone 1

Chloritoid-chlorite-kyanite forms the most northern and lowest grade zone. The 'maximum' assemblages of this zone contain only three 'coexisting' phases in the system $\text{FeO-MgO-Al}_2\text{O}_3$ ($-\text{SiO}_2-\text{K}_2\text{O}-\text{H}_2\text{O}$) – chloritoid, chlorite and kyanite.

Restricted exposure prevented the collection of more than a few samples from this zone. Of the 22 that were obtained (21 from the top of the pass road and one from northern Val di Campo), 7 bore the defining assemblage.

The northern limit of zone 1, not shown in Figure 2, may be assumed to coincide with E. Niggli's (1970) Alpine kyanite 'isograd', which runs essentially E-W through the peak of Scopi. The first appearance of staurolite marks the southern, high grade boundary.

3b. Zone 2

Staurolite-chloritoid-chlorite-kyanite. The first tiny staurolites in the area are found just E of the highest point on the pass road and hence $1\frac{1}{2}$ km (topographically) North of E. Niggli's (1970) staurolite 'isograd'. Their appearance marks the boundary of a second mineral zone in which 'maximum' assemblages bearing chlorite, chloritoid, staurolite and kyanite are common. The boundary is not marked by the concomitant loss of one of the zone 1 phases.

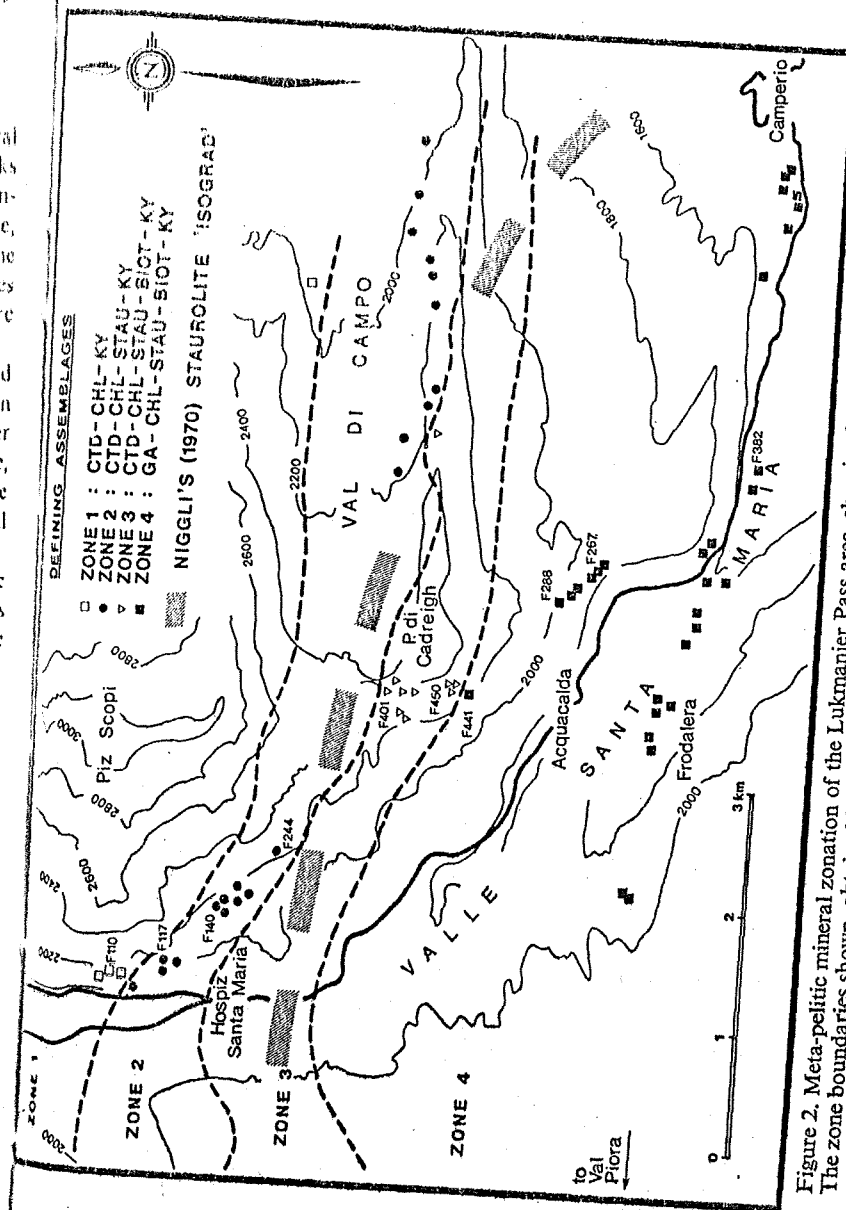


Figure 2. Meta-pelitic mineral zonation of the Lukmanier Pass area, showing locations of 'maximum' assemblages. The zone boundaries shown, obtained by the superposition of structural contours on the Lukmanier topography, represent a dip of 35° N (see text).

As in the other mineral zones mapped, there is a significant paucity of non-defining assemblages here as well as a complete lack of any trend towards higher variance assemblages southwards. In fact, out of 90 zone 2 samples sectioned, 46 were found to be defining ones. Non-defining samples generally lack kyanite and, at low grade, staurolite.

3.c. Zone 3

Biotite-staurolite-chloritoid-chlorite-kyanite is ushered in by the appearance in maximum assemblages of biotite; the new phase appears without the concurrent loss of any pre-existing one. The first plagioclase in the basal Stgir is also found at the zone 3 lower boundary, again simply as an addition.

3.d. Zone 4

Biotite-staurolite-garnet-chlorite-kyanite is the highest grade and most extensive zone in the area – samples bearing the defining 5-phase assemblage (chloritoid is present only as abundant inclusions in garnet) are common throughout the southern half of the pass. The first 'maximum' assemblage was found at the southernmost tip of Piz di Cadreigh (a sample bearing all six Fe-Mg-Al silicates was collected several m N of the first five-phase sample), where the low grade boundary of the zone was drawn.

From petrographic analysis of some probable basal Stgir samples from Val Piora it is evident that the zone 4 defining assemblages occur over much of the area to the W of the pass. There is some evidence that it is still stable (albeit in Pennide rather than Ultrahelvetic schists) in the Piz Molare area 5 km S of Valle Santa Maria (Thakur, 1971, unpubl. thesis, London Univ.). The high grade boundary of this zone must therefore lie well to the S of the Lukmanier area.

A zone boundary trace for the zone 2–zone 3 transition was obtained by drawing structural contours on the given topography on a trial and error basis and assuming a planar surface with various northward dips upwards from zero, until a boundary was generated that fitted the data. Trial and error was also necessary to produce the boundary strike direction shown, but here the E–W trend of E. Niggli's (1970) staurolite 'isograd' was available as a rough guide. The zone 2–zone 3 boundary in Figure 2 assumes a dip of 35° N, a WNW–ESE strike and uses the one known boundary exposure in Val di Campo as a control point. This value of 35° must be considered a minimum, however, since greater dips also satisfy the observations. It is in agreement with that quoted by Wunderlich (1957) for 'isograds' in the region generally.

In so far as the zone boundary reactions deduced have steep, near-parallel slopes in P–T space (see Fig. 3) it was thought reasonable to draw the other, less well controlled transitions on the basis of the 35° minimum. The traces of these boundaries are also shown in Figure 2.

Complete assemblages (minus apatite, zircon, tourmaline and graphite) of ten defining samples from the four Lukmanier mineral zones are listed in Table 1. Mean grain-edge microprobe analyses of the phases in 3 of the samples are given in Tables 2–4; details of the analytical procedure are given in Fox (1974).

4. Zone boundary reactions

It is clear that much of the basal Stgir contains an over-abundance of apparently coexistent phases. Even when all 11 major and minor J-components (MgO, FeO, Fe₂O₃, Al₂O₃, ZnO, K₂O, Na₂O, SiO₂, TiO₂, MnO and CaO) are considered, the 12- or 13-phase assemblages typical of zones 3 and 4 remain univariant or invariant – an unlikely situation in phase rule terms.

Table 1. Complete thin-section-scale mineralogy of some selected Stgir meta-pelites showing 'maximum' assemblages. Numbers in parentheses correspond to the Harker Petrology collection of Cambridge University

Zone 1	
F-110 (111327)	chlorite-chloritoid-kyanite-muscovite-paragonite-clinozoisite/epidote-quartz-rutile-ilmenite
Zone 2	
F-117 (111334)	chlorite-chloritoid-staurolite-kyanite-muscovite-paragonite-clinozoisite/epidote-quartz-rutile-ilmenite
F-140 (111349)	chlorite-chloritoid-staurolite-kyanite-muscovite-paragonite-quartz-rutile-ilmenite
F-244 (111411)	chlorite-chloritoid-staurolite-kyanite-muscovite-paragonite-clinozoisite/epidote-quartz-rutile-ilmenite
Zone 3	
F-401 (111500)	chlorite-biotite-chloritoid-staurolite-kyanite-muscovite-margarite-clinozoisite/epidote-quartz-rutile-ilmenite
F-450 (111538)	chlorite-biotite-chloritoid-staurolite-kyanite-muscovite-paragonite-clinozoisite/epidote-plagioclase-quartz-rutile ilmenite
Zone 4	
F 441 (111530)	chlorite-biotite-staurolite-kyanite-garnet-muscovite-paragonite-clinozoisite/epidote-plagioclase-ilmenite-quartz
F-288 (111437)	chlorite-biotite-staurolite-kyanite-garnet-muscovite-paragonite-clinozoisite/epidote-plagioclase-ilmenite-quartz
F-267 (111424)	chlorite-biotite-staurolite-kyanite-garnet-muscovite-margarite-clinozoisite/epidote-plagioclase-ilmenite-quartz
F-382 (111489)	chlorite-biotite-staurolite-kyanite-garnet-muscovite-paragonite-clinozoisite/epidote-plagioclase-ilmenite-rutile-quartz

Simple analogies (e.g. the binary solvus of the magnetite-hematite oxygen buffer) suggest that solid substitutional behaviour where a reacting fluid component (such as H₂O) is internally buffered should be regular with increasing grade. Under certain conditions, no compositional changes at all would be observed (cf. Evans & Guidotti, 1966; Guidotti, 1970). Significant irregularities superimposed upon a regular, prograde, Fe/Mg trend in the minerals of the basal Stgir (Fox, 1974) thus suggests the presence during metamorphism of anomalies in P_{H₂O}, presumably reflecting spasmodic influxes of unreactive CO₂ from adjacent Stgir or Quarten marls. This is consistent only with the external control of P_{H₂O}, so that no solution to the observed phase rule violation can be sought from this quarter. Detailed microprobe work has shown that most of the phases in these rocks are chemically inhomogeneous to a surprising degree both within individual grains and from grain to grain in a single thin section. This mineral-

Table 2. Mean edge compositions of F-140 minerals (Zone 2). Standard Deviations in parentheses

	Staurolite	Chloritoid	Chlorite	Ilmenite	Muscovite	Paragonite
SiO ₂	28.49 (0.19)	24.30 (0.16)	25.40 (0.16)	00.25 (0.03)	46.24 (0.46)	45.52 (0.74)
TiO ₂	00.54 (0.02)	00.04 (0.02)	00.09 (0.02)	52.60 (0.15)	—	—
Al ₂ O ₃	54.57 (0.32)	40.97 (0.55)	23.50 (0.38)	00.18 (0.03)	35.62 (0.42)	39.47 (0.39)
FeO*	11.88 (0.19)	22.25 (0.17)	21.81 (0.33)	46.55 (0.21)	00.85 (0.03)	00.16 (0.07)
ZnO	00.45 (0.02)	00.06 (0.02)	00.06 (0.02)	—	—	—
MnO	00.09 (0.01)	00.17 (0.02)	00.08 (0.01)	00.38 (0.02)	—	—
MgO	01.58 (0.09)	04.08 (0.20)	16.91 (0.25)	00.07 (0.02)	00.80 (0.04)	00.15 (0.05)
CaO	—	—	—	—	00.02 (0.00)	00.81 (0.07)
Na ₂ O	—	—	—	—	01.60 (0.10)	06.16 (0.19)
K ₂ O	—	—	—	—	08.50 (0.90)	01.43 (0.18)
Anhydrous total	97.60	91.87	86.95	100.05	93.63	93.90
Si	7.849	1.998	5.112	0.013	6.190	5.922
Al	—	—	2.888	—	1.810	2.078
Al	17.722	3.971	2.890	0.011	3.812	3.975
Ti	0.113	0.003	0.014	1.990	—	—
Fe	2.738	1.530	3.806	1.959	0.096	0.040
Zn	0.091	0.004	0.010	—	—	—
Mn	0.022	0.012	0.014	0.017	—	—
Mg	0.647	0.500	5.256	0.006	0.159	0.029
Ca	—	—	—	—	0.003	0.114
K	—	—	—	—	0.415	1.555
Mn	—	—	—	—	1.452	0.238
O	46.000	12.000	28.000	6.000	22.000	22.000

* Total iron calculated as FeO.

chemical inhomogeneity may be a function of the graphitic nature of the basal Stgir and/or the proximity of calcareous rocks and hence the absolute concentration of H₂O in the fluid. Thus, mosaic equilibrium, or 'disequilibrium' (the subject of a paper in preparation) must be almost entirely responsible for the apparent low variance.

'Disequilibrium' on such an intimate scale renders impracticable the identification of detailed boundary reactions or rigidly categorized ion exchanges. Nevertheless, the sharply defined nature of the mineral zones strongly suggests that, on a broad scale, a small number of mineral transformations are to be recognized. The sampling horizon is essentially isochemical and thus if the influence of minor components is excepted, it seems reasonable to infer that the operative reactions were discontinuous and that the zone boundaries approximate to isograds in the sense of Carmichael (1970).

Even though 'disequilibrium' is prevalent, sequences of thin sections through the mineral zones show rough modal increases in product and concomitant decreases in the reactant species of the inferred reactions. Tie lines allowed by these reactions, and particularly 3-phase areas, show similar rough increases in abundance with grade relative to those not allowed.

In zone 2, for example, a decrease in chloritoid modality is strongly tied to an increase in that of staurolite. Chlorite abundance is weakly tied to that of staurolite, but the sign of the kyanite modality change is problematical. From what a

Table 3. Mean edge compositions of F-450 minerals (Zone 3). Standard deviations in parentheses

	Staurolite	Chloritoid	Biotite	Chlorite
SiO ₂	27.68 (0.20)	24.20 (0.17)	36.08 (0.25)	24.35 (0.12)
TiO ₂	00.43 (0.02)	00.00 (0.00)	01.59 (0.04)	00.07 (0.03)
Al ₂ O ₃	54.78 (0.44)	40.97 (0.38)	19.06 (0.38)	22.81 (0.25)
FeO	13.50 (0.17)	22.45 (0.16)	16.91 (0.61)	22.57 (0.09)
ZnO	00.09 (0.02)	00.03 (0.02)	00.02 (0.02)	00.02 (0.02)
MnO	00.16 (0.01)	00.20 (0.02)	00.03 (0.00)	00.09 (0.01)
MgO	01.67 (0.08)	03.99 (0.10)	11.75 (0.43)	16.61 (0.12)
CaO	—	—	00.02 (0.02)	—
K ₂ O	—	—	08.89 (0.06)	—
Anhydrous total	98.31	91.83	94.35	86.52
Si	7.630	1.992	5.451	5.132
Al	17.806	3.975	2.549	2.868
Al	—	—	0.845	2.799
Ti	0.090	0.000	0.181	0.012
Fe	3.112	1.546	2.137	3.978
Zn	0.019	0.002	0.002	0.003
Mn	0.037	0.019	0.004	0.017
Mg	0.687	0.490	2.647	5.219
Ca	—	—	0.003	—
K	—	—	1.713	—
O	46.000	12.000	22.000	28.000

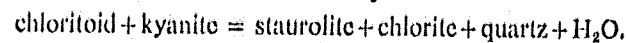
	Ilmenite	Muscovite	Paragonite	Plagioclase
SiO ₂	00.20 (0.01)	46.34 (0.29)	44.15 (0.48)	59.59 (0.42)
TiO ₂	51.95 (0.60)	—	—	—
Al ₂ O ₃	00.08 (0.01)	35.49 (0.16)	41.26 (0.09)	25.58 (0.16)
FeO	46.52 (0.19)	01.19 (0.03)	00.36 (0.01)	—
ZnO	—	—	—	—
MnO	00.51 (0.02)	—	—	—
MgO	00.24 (0.12)	00.97 (0.02)	00.13 (0.02)	—
CaO	—	00.02 (0.01)	01.46 (0.14)	06.84 (0.22)
Na ₂ O	—	01.62 (0.07)	06.57 (0.11)	07.82 (0.12)
K ₂ O	—	08.64 (0.17)	00.85 (0.05)	00.07 (0.02)
Anhydrous total	99.50	94.27	94.78	99.90
Si	0.010	6.179	5.706	10.631
Al	0.005	1.821	2.294	5.381
Al	—	3.760	3.992	—
Ti	1.981	—	—	—
Fe	1.973	0.133	0.040	—
Zn	—	—	—	—
Mn	0.022	—	—	—
Mg	0.019	0.194	0.026	—
Ca	—	0.003	0.203	—
Na	—	0.418	1.645	1.307
K	—	1.471	0.140	2.704
O	6.000	22.000	22.000	32.000

Table 4. Mean edge compositions of F-288 minerals (Zone 4). Standard deviations in parentheses

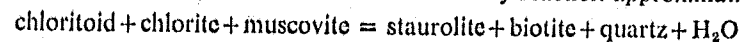
	Garnet	Staurolite	Biotite	Chlorite
SiO ₂	37.12 (0.27)	27.86 (0.41)	36.25 (0.22)	24.60 (0.09)
TiO ₂	00.07 (0.02)	00.45 (0.03)	01.43 (0.05)	00.18 (0.04)
Al ₂ O ₃	21.51 (0.66)	54.02 (0.56)	19.53 (0.43)	23.42 (0.65)
FeO	33.71 (0.69)	13.20 (0.12)	15.51 (0.55)	20.39 (0.41)
ZnO	00.00 (0.00)	00.03 (0.02)	00.05 (0.02)	00.02 (0.02)
MnO	01.22 (0.03)	00.08 (0.01)	00.05 (0.01)	00.08 (0.02)
MgO	02.88 (0.08)	01.04 (0.10)	11.73 (0.38)	16.98 (0.16)
CaO	04.17 (0.24)	—	00.00 (0.00)	—
K ₂ O	—	—	08.97 (0.06)	—
Anhydrous total	100.68	97.55	93.52	85.59
Si	5.926	7.722	5.484	5.167
Al	0.074	17.655	2.516	2.833
Al	3.974	—	0.967	2.967
Ti	0.009	0.094	0.163	0.028
Fe	4.500	3.061	1.962	3.581
Zn	0.000	0.007	0.000	0.003
Mn	0.166	0.019	0.006	0.014
Mg	0.685	0.803	2.644	5.316
Ca	0.714	—	0.000	—
K	—	—	1.732	—
O	24.000	46.000	22.000	28.000

	Ilmenite	Muscovite	Paragonite	Plagioclase
SiO ₂	00.15 (0.05)	46.84 (0.26)	44.62 (0.49)	61.65 (0.61)
TiO ₂	53.30 (0.28)	—	—	—
Al ₂ O ₃	00.13 (0.02)	36.09 (0.41)	40.61 (0.31)	24.01 (0.27)
FeO	46.19 (0.27)	00.79 (0.02)	00.30 (0.03)	00.14 (0.01)
ZnO	—	—	—	—
MnO	00.29 (0.01)	—	—	—
MgO	00.14 (0.01)	00.76 (0.18)	00.12 (0.01)	00.03 (0.02)
CaO	—	00.02 (0.02)	01.16 (0.06)	05.19 (0.26)
Na ₂ O	—	01.80 (0.16)	05.62 (0.14)	08.68 (0.17)
K ₂ O	—	08.54 (0.10)	00.98 (0.07)	00.07 (0.01)
Anhydrous total	100.20	94.84	94.31	99.77
Si	0.008	6.190	5.786	10.963
Al	0.008	1.810	2.214	5.034
Al	—	3.812	3.995	—
Ti	2.008	—	—	—
Fe	1.936	0.088	0.033	0.022
Zn	—	—	—	—
Mn	0.013	—	—	—
Mg	0.011	0.150	0.023	0.009
Ca	—	0.003	0.162	0.990
Na	—	0.462	1.639	2.993
K	—	1.441	0.163	0.017
O	6.000	22.000	22.000	32.000

known about Fe-Mg partitioning between the various phases (e.g. Albee, 1972) and from the chemography suggested for the zone 2 specimens by the microprobe analyses in Table 3, it is reasonable to assume the following to have been approximated at the zone 1-zone 2 boundary:

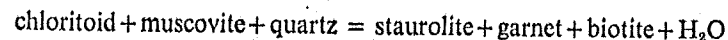


The reaction responsible for the zone 3 mineralogy must reconcile the appearance of biotite with an observed increase in staurolite and a decrease in chloritoid and chlorite modalities. Kyanite abundance again does not seem to vary in any regular way. A zone 2-zone 3 boundary reaction approximating to



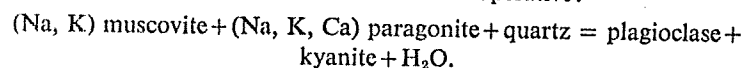
must have been operative.

Garnet first appears and matrix chloritoid disappears at the zone 3-zone 4 boundary. From this boundary southwards biotite shows a strong modal increase while a smaller increase is noticed for staurolite. The reaction



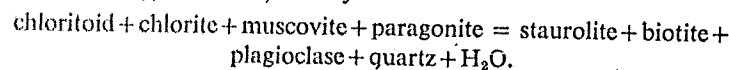
is most consistent with these observations. The cause of an observed moderate decrease in the abundance of both chlorite and kyanite here is problematical.

The often very fine white-mica intergrowths and the skeletal, amoeboid character of plagioclase renders impractical modal analysis of the precision required to decipher the role of these phases in the zone boundary reactions. Knowing that plagioclase abruptly appears and increases in abundance from the zone 2-zone 3 boundary southwards, however, some surmise does seem possible. Thus it can be assumed that the production of feldspar in the sampling horizon was a consequence of the first appearance of biotite. The requisite breakdown of muscovite must have liberated significant Na₂O (zone 2 muscovites have molecular compositions averaging Mu₆₄ Parag₂₂ Mg-celad₃ Fe-Celad₆) and because the CaO in the resulting plagioclase could only have been made up by paragonite (plagioclase compositions fall in the range oligoclase-andesine while zone 2 paragonites average Mu₅ Parag₈₀ Marg₈ Mg-Celad₁ Fe-Celad₂) the following must have been the fundamental transformation operative:

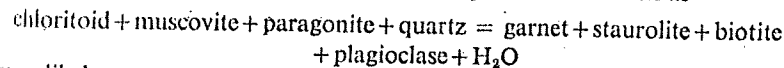


A clarification of the role of rare margarite in this and other Lukmanier reactions must await further study.

It may be that the above was coupled to the ferromagnesian boundary reaction not only through biotite, but also by the contribution of the excess alumina not to kyanite but to, say, staurolite, namely



The behaviour of muscovite and paragonite in the zone 3-zone 4 boundary reaction must have been similar to that above, since plagioclase modality appears to continue to increase with grade. Thus a coupled reaction such as



seems likely.

5. The petrogenetic grid and the conditions of metamorphism

The basic information about the zone boundary reactions obtained above can be used to fit the Lukmanier parageneses into a more general scheme. Two Schreinmakers constructions, taken essentially from Albee (1965) and Hoschek (1969), but differing in that the non-collinear chemographic relationship between Al_2SiO_5 , staurolite, chloritoid and garnet obtained from Tables 2-4 is assumed, are shown in Figure 3. The grids differ from Albee's in that equilibria involving all three Al_2SiO_5 polymorphs are considered and they differ from Hoschek's in that the invariant point [Ky, sill, and cd] is taken to lie in the kyanite field, in accordance with observations made here and elsewhere (Fox, 1971). It should be noted that Hoschek's location of the cordierite-limiting reaction at the low P part of Figure 3(a) has been confirmed by Hess (1969) and by Bird & Fawcett (1973), who limit even Mg-cordierite to the vicinity of the Al_2SiO_5 triple point as determined by Richardson, Gilbert & Bell (1969). Figure 3(b), which probably corresponds to the condition $P_{\text{H}_2\text{O}} \approx 0.5P_{\text{total}}$ (cf. Hoschek, 1969), was generated from Figure 3(a) by rotating the key equilibria - (ky, and, sill, stau, cd), (ky, and, sill, ga, cd) and (ky, and, sill, chl, cd) - about their respective low-temperature invariant points by roughly the amount prescribed for the analogous equilibria by Albee (1965) and by assuming the upper stability limit of cordierite to be relatively insensitive to changes in $P_{\text{H}_2\text{O}}$.

As observed by Albee (1972), the same sequence of reactions is encountered in the vicinity of the invariant point [ky, and, sill, cd] - the area of P-T space corresponding to the Lukmanier situation - whether the situation is that of Figure 3(a) or that of Figure 3(b). It is therefore impossible to determine purely on paragenetic grounds which of these most closely corresponds to the Lukmanier situation. With a positive sloping geothermal gradient emanating from the shaded regions in the two grids it seems unlikely that cordierite-bearing assemblages will be encountered at high grade, irrespective of fluid composition. This observation can be used to compare the Lepontine metamorphism (represented by such a gradient) with the problematical rocks around the Bergell granite. Many geologists (e.g. E. Wenk, 1970; Moticska, 1970) consider the Bergell event to represent a high level continuation of the Lepontine phase. However, such paragenesis as that of the cordierite-sillimanite-garnet-biotite contact hornfels reported by E. Wenk (1973) from the latter area must be incompatible with the Lepontine gradient and must lie in the vicinity of (stau, ky, and, chl, ctd), far removed from the relevant bundle about [ky, and, sill, cd]. They, and hence the whole Bergell event, must require an entirely new geothermal path.

The indifferent crossing of the equilibria involving chloritoid + kyanite (Hoschek, 1967; Richardson, 1968) and margarite + quartz (Storre & Nitsch, 1974) provide an estimate for the conditions of metamorphism here. The two equilibria intersect at 5.5 kb and 550 °C when $P_{\text{H}_2\text{O}} = P_{\text{total}}$. (Margarite has been microchemically identified in a number of carbonate-free, 'low variance' basal Stgir rocks from zones 2, 3, and 4 (Fox, 1974). The presence of graphite suggests that $P_{\text{H}_2\text{O}}$ in the basal Stgir is unlikely to have exceeded $0.85 P_{\text{total}}$ (French, 1966) and since a lowering of $P_{\text{H}_2\text{O}}$ depresses the temperatures of both decompositions to a significant extent these figures must be considered maximum estimates. They are,

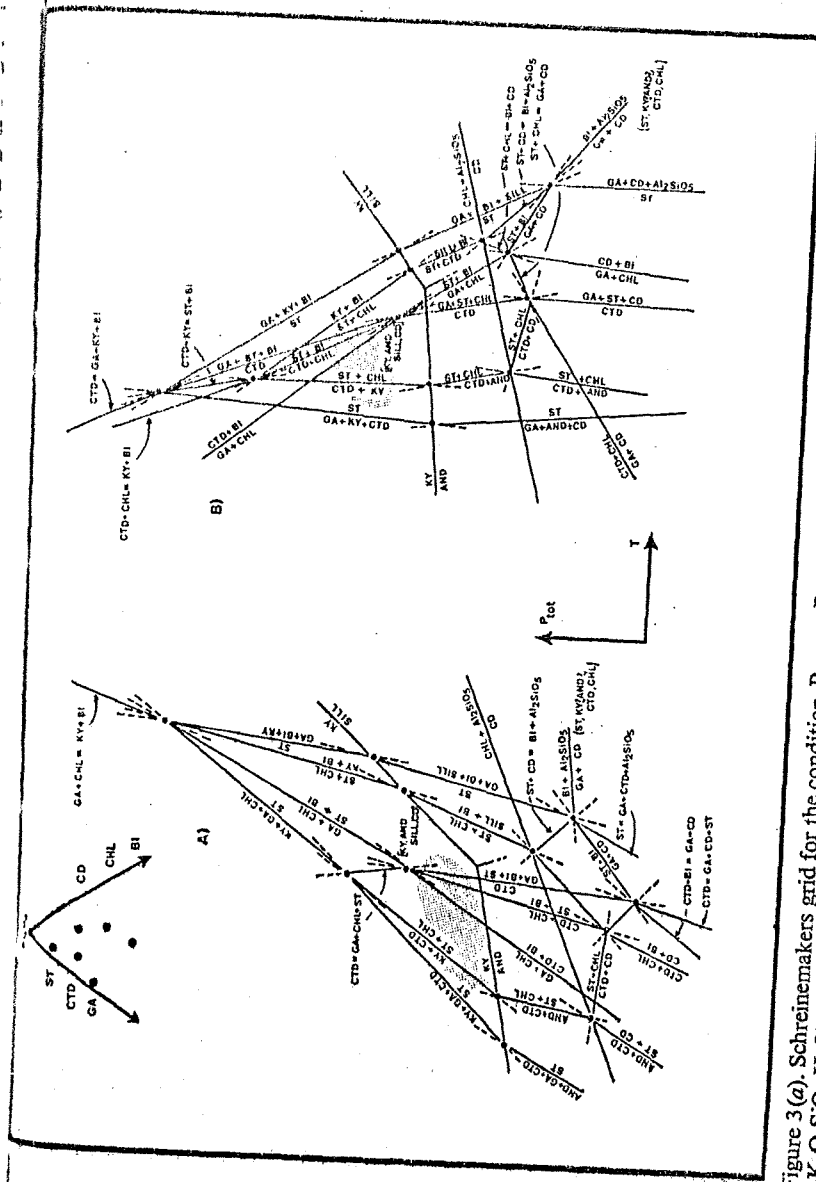


Figure 3 (a). Schreinmakers grid for the condition $P_{\text{H}_2\text{O}} = P_{\text{total}}$ involving nine phases in the system $\text{FeO-MgO-Al}_2\text{O}_3$ ($\text{-K}_2\text{O-SiO}_2\text{-H}_2\text{O}$) and assuming the phase chemography shown inset. Not all possible equilibria included. Quartz and muscovite are in excess. (b) Schreinmakers grid for the condition $P_{\text{H}_2\text{O}} P_{\text{total}}$ derived from (a) (see text). The relation between several invariant points and the andalusite-sillimanite inversion is ambiguous; thus the equilibrium and = sill has not been extended into the low P-high T portions.

however, in good agreement with those obtained by Frey (1969). When linearly extrapolated to zero these values give an approximate figure of 18–24 °C/km for the operative thermal gradient, which is comparable with current steady state shield values (Dewey & Pankhurst, 1970), but which is admittedly at odds with the convective heat source proposed below.

6. Discussion

Sampling in the pass has thus enabled the delineation of three-dimensional zone boundaries which may be characterized by simple, discontinuous reactions. Since these reactions have relatively steep slopes regardless of fluid composition (i.e. are relatively insensitive to changes in total pressure, see Fig. 3) it follows that these boundaries can be considered as paleoisotherms.

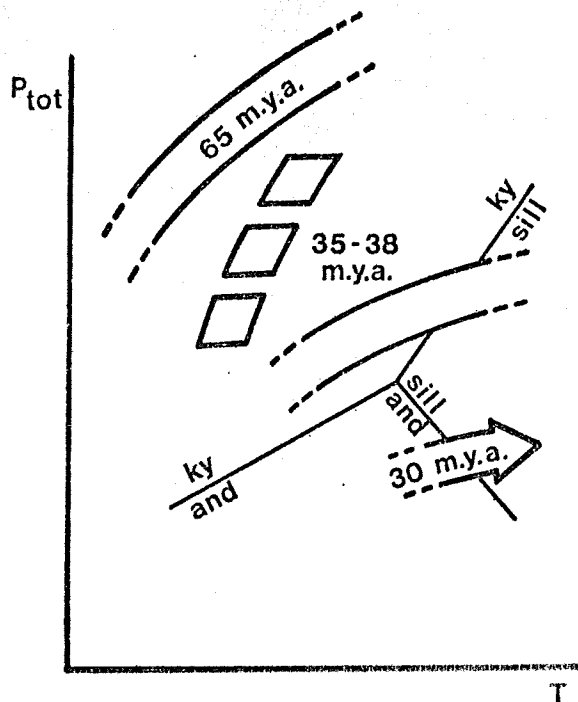


Figure 4. Assumed P-T trajectories and age relations of the various stages of Swiss Alpine metamorphism. The dashed curve represents the stage of prasinization discussed by Bearth (1967) and Ernst (1973).

Data is now available from heat flow and geochronological calculations (Clark & Jäger, 1969) and from railway surveys (Schaer & Jeanrichard, 1974) which clearly indicate a late- and post-orogenic differential unloading of the mountain chain centred about the southern Lepontine. When it is assumed that 35 Ma have passed since the peak of metamorphism (Hunziker, 1970) an estimate of total uplift can be derived from the data: in particular for the Lukmanier area, as

elevation of some 28 km. A moment of rotation about the Eocene topographic contours along the section Lucerne–Lukmanier is suggested which, when translated into real values relative to the direction of increasing grade works out to 14–21°. Note, incidentally, that this must certainly be an over-estimation since Schaer & Jeanrichard's data force the assumption that the Lucerne area (which today is marked by almost 2 km of elevation) has suffered negligible uplift.

Bearing in mind the minimum true dip of 35°N for the zone 2–zone 3 boundary (hence, isotherm) and the post-orogenic rotation of the Eocene topography, a true minimum value of 14–21°N for the inclination of the Lukmanier paleo-isotherms can be obtained. This, in turn, can be applied to the problems of the source of heat for the Lepontine metamorphic phase. A situation wherein isotherms were significantly prograde inclined does not conform to the thermal model for oceanic trenches (Oxburgh & Turcotte, 1971; Sugimura & Uyeda, 1973) in which blueschist tectonites are thought to form. From a mechanistic point of view, therefore, this seems to confirm that the Lepontine gradient cannot be a high-temperature continuation or erosional derivative of the Alpine blueschists. It is also clear that the horizontal isotherms required by E. Niggli's (1970) burial hypotheses for the Lepontine did not obtain here. Although burial metamorphic isotherms need not be perfectly horizontal, the true minimum of 14–21° is several times larger than thermal perturbations which might be induced, for example, by mantle-crust angular irregularities (see O'Brien, 1968). Clearly, some thermal convection located S of the Lukmanier area is required.

Further isograd mapping may elucidate the nature of this convective source and its relation to the distinct blueschist and Bergell events (see Frey *et al.* (1974) for a review of the Bergell problem), but one line of thought so far inadequately considered may be worth following up. Both the Lepontine and Bergell events may represent a single, drawn-out stage of magmatic convection subsequent to lithospheric subduction. The observed geothermal differences between the two may represent only the degree to which late-orogenic transcurrent faulting along the Insubric line and its offshoots allowed the melt to concentrate near the earth's surface.

7. Conclusions

The somewhat subjective study of the mineral zone transitions – the best that can be done in view of the ubiquity of 'disequilibrium' in the Lukmanier metapelites – yields some information of relevance to the larger problem of Alpine orogenesis. It now seems reasonable to assume different thermal regimes for different parts of the Alpine belt not only because gradients seem to differ (10–20 °C/km in the perimetal blueschists, from Chinner & Dixon, 1973; 18–24 °C/km for the Lukmanier section of the Lepontine; 30 °C/km, i.e. below the Al_2SiO_5 triple point, for the Bergell contact aureole) but also because P-T conditions in the various parts are such that no reasonable, smooth gradient can be drawn between them. It must, therefore, be concluded that the Alpine orogeny was thermally plural, that at least 3 distinct prograde phases can be recognized (a fourth, pre-blueschist, high T-low P phase has been reported in the Eastern Alps by Hawkesworth, Waters & Bickle, 1975), and that the intermediate,

Lepontine phase had a convective heat source. This interpretation (excepting the nature of the heat source) is essentially in agreement with that of E. Niggli (1970) and Zwart (1974) and is, in fact, borne out by the most recent geochronological findings (Jäger, 1973; Gulson, 1973). Figure 4 shows the proposed P-T trajectories for the major metamorphic phases and illustrates this remarkable tendency to higher crustal levels of metamorphism with decreasing age.

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Marsupiocrinidae of the Laurel Limestone of southeastern Indiana

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(Plate 1)

Summary. Three new species of *Marsupiocrinus* are described from the basal Wenlock Laurel Limestone of southeastern Indiana. Although the earliest representatives of the genus yet known, they display fully the specialized characters peculiar to it. Two new subgenera of *Marsupiocrinus* are erected on the basis of arm structure and cup shape.

1. Introduction

Bassler & Moodey (1943, pp. 44-45) list a sizeable echinoderm fauna, including 59 crinoids, from the Laurel Limestone (Laurel member, Salamonie Dolomite of Pinsak & Shaver (1964)) at St Paul, Indiana. Most of these were described by Miller (1892, 1894) and by Springer (1926). Since the publication of Springer's monograph little additional research has been done on the Laurel echinoderms. The discovery of a prolific Laurel locality near Sandusky, Indiana (Frest & Paul, 1971), the rediscovery of Springer's St Paul locality, and the examination of the old Laurel collections have turned up a number of new echinoderms, among which are the crinoids described below.

Most of Springer's and Miller's material, including the specimens of *Marsupiocrinus*, was collected from the Adams' Quarry, SE of St Paul (SW $\frac{1}{4}$ SW $\frac{1}{4}$ SE $\frac{1}{4}$ sec. 3, T11N R8E, Adams' 7 $\frac{1}{2}$ ' quadrangle, Decatur County, Indiana). This is the more southerly of the two quarries mentioned by Foerste (1898, pp. 230-1) as being owned by H. C. Adams. It is still known locally as the Adams' Quarry, and is the only quarry in the area which was worked regularly during the period 1892-1926. The other quarries described by Foerste can still be located; these proved to be almost totally unfossiliferous. A stratigraphic section of the Adams' Quarry is given below (Fig. 1): it differs largely from that listed by Foerste (1898) since the operation was expanded considerably between 1921-25 before closing permanently.

2. Systematic descriptions

- Class Crinoidea Miller, 1821
- Subclass Camerata Wachsmuth & Springer, 1886
- Order Monobathrida Moore & Laudon, 1943
- Suborder Glyptocrinina Moore, 1952
- Superfamily Patelliocrinacea Ubaghs, 1953
- Family Marsupiocrinidae Jaekel, 1918

Definition. Monocyclic camerates with low cup and flat or nearly flat base.

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