二十世纪中国女性文学史

ROCK-FORMING MINERALS

Vol. 1   Ortho- and Ring Silicates

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CONTENTS *Vol. 1. Ortho- and Ring Silicates*

OLIVINE GROUP

The olivine group includes a number of closely related minerals which crystallize with orthorhombic symmetry. The structures of all the minerals of the group consist of independent $[SiO_4]$ tetrahedra linked by divalent atoms in six-fold coordination. As is common in other orthosilicate minerals silicon is not replaced by aluminium, and the octahedral positions in the structures are occupied almost exclusively by divalent atoms; the common trivalent atoms Al and Fe^{+3} are either absent, or present in very small amounts. In the (Mg,Fe)-olivines there is a continuous series between the two end-members Mg_2SiO_4 (forsterite) and Fe_2SiO_4 (fayalite): the fayalite–knebelite ($FeMnSiO_4$)–tephroite (Mn_2SiO_4) series also shows complete solid solution. In the forsterite–fayalite series, however, manganese occurs only in small amounts, and the low content of manganese in this series is probably related to the small amounts of manganese in the magmas and rocks in which they crystallized. A few occurrences of magnesium-rich tephroite, picrotephroite, have been reported. The $CaMgSiO_4$ orthosilicate (monticellite) does not vary greatly from the ideal composition, and members of the forsterite–fayalite series in general do not contain appreciable amounts of calcium. The iron analogue of monticellite, kirschsteinite, $CaFeSiO_4$, is known from slags, but has not been reported from a natural occurrence, and a magnesium-rich kirschsteinite containing 69 per cent. $CaFeSiO_4$ is the most iron-rich mineral of the Fe_2SiO_4–$CaFeSiO_4$ series yet reported. Glaucochroite ($CaMnSiO_4$) is a rare olivine, and minerals intermediate in composition between $CaMgSiO_4$ and $CaMnSiO_4$ are unknown. Larsenite ($PbZnSiO_4$) has normally been included in the olivine group; a recent determination of the cell size and space group, however, has shown that it is not isostructural with the olivine group. Single-crystal X-ray investigation of larsenite from Franklin, New Jersey, gave a 8·23 Å, b 18·94 Å, c 5·06 Å (a and b are approximately double the values for olivine), space group *Pnam* or *Pna*, $Z = 8$ (Layman, 1957). The minerals of the olivine group are very susceptible to alteration by hydrothermal and weathering processes and are often pseudomorphed by minerals of low temperature paragenesis.

The (Mg,Fe)-olivines are common and important rock-forming minerals, and are particularly characteristic of the basic and ultrabasic igneous rocks; often their composition is a useful indication of the differentiation stage of the parent magmas in which they crystallized. Monticellite is not a common mineral but occurs in some ultrabasic rocks and in limestone contacts with basic and acid intrusions. The other olivine minerals are rare and have a very restricted paragenesis.
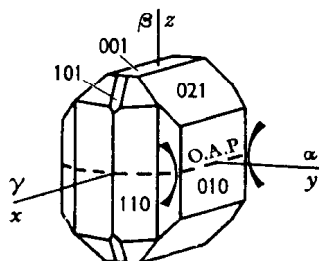
Olivine
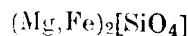
$(Mg, Fe)_2[SiO_4]$

ORTHORHOMBIC $(+)(-)$

Forsterite	Fayalite
Mg_2SiO_4	Fe_2SiO_4

	Forsterite	Fayalite
α^*	1·635	1·827
β	1·651	1·869
γ	1·670	1·879
δ	0·035	0·052
$2V_\gamma$	82°	134°
	$\alpha=y, \beta=z, \gamma=x$, O.A.P. (001).	$\alpha=y, \beta=z, \gamma=x$, O.A.P. (001).
Dispersion:	$r>v$	$r>v$
D†	3·222	4·392
H	7	$6\frac{1}{2}$
Cleavage:	{010},{100}imperfect.	{010}moderate, {100}weak.
Twinning:	{100},{011},{012}	{100}
Colour:	Green, lemon-yellow; colour-less in thin section.	Greenish yellow, yellow-amber; pale yellow in thin section.
Pleochroism:	—	$\alpha=\gamma$ pale yellow
		β orange yellow
Unit cell:†	a 4·756 Å	a 4·817 Å
	b 10·195 Å	b 10·477 Å
	c 5·981 Å	c 6·105 Å
	Z=4	Space group *Pbnm*
	Gelatinizes in HCl,	

* Values of refractive indices, birefringence and 2V refer to end-members, between which there is continuous variation.

† Values for synthetic forsterite and fayalite (Yoder & Sahama, 1957).

The magnesium-rich olivines are common constituents of ultrabasic and basic rocks. The iron-rich members crystallize in small quantities in some acid and alkaline plutonic and volcanic rocks. Magnesium-rich olivines also occur in thermally metamorphosed impure dolomitic limestones, *e.g.* forsterite marble. The iron-rich varieties occur as a product of regional metamorphism in eulysites and other metamorphosed iron-rich sediments. Although the two atomic pairs $MgFe^{+2}$ and $Fe^{+2}Mn$ show complete diadochy, the (Mg,Fe)-olivines rarely contain appreciable amounts of Mn. The Mg-olivine, forsterite, is named after J. Forster, founder of the Heuland Cabinet, and the Fe-olivine, fayalite, after Fayal Island in the Azores where it was believed to have occurred in a local volcanic rock, but it was probably obtained from slag carried as ship's ballast (Palache, 1950). The names forsterite and fayalite are restricted to the compositions Fo_{100-90} and Fo_{10-0} respectively; the nomenclature of the intermediate members of the series is given in Fig. 11. The clear gem quality olivine is known as peridot.

STRUCTURE

The olivine structure was determined by Bragg and Brown (1926) on a mineral of composition $Fo_{90}Fa_{10}$, the cell dimensions of which are a 4·755 Å, b 10·21 Å, c 5·985 Å. The structure consists of individual silicon–oxygen tetrahedra linked by magnesium atoms each of which has six nearest oxygen neighbours. The oxygens lie in sheets parallel to the (100) plane and are
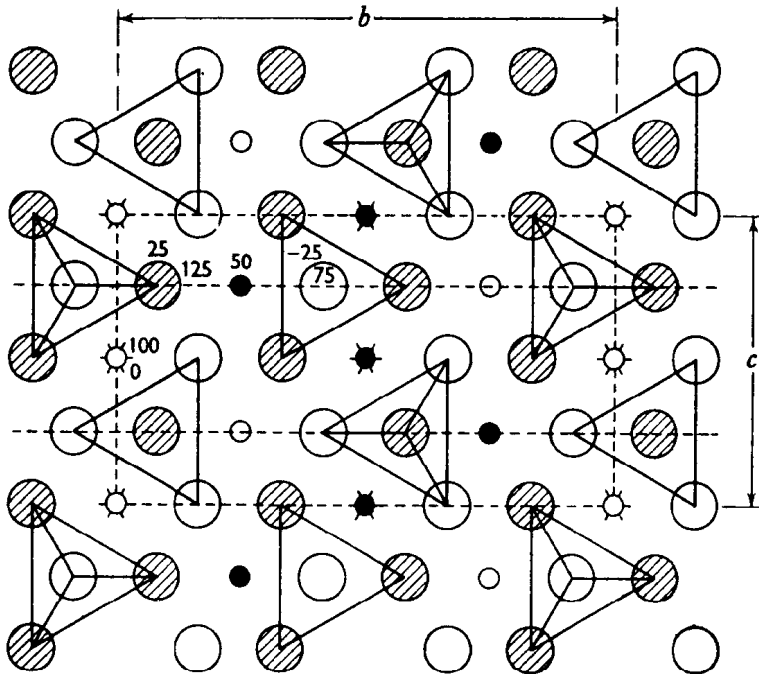

Fig. 1. Olivine structure parallel to (100) plane. Si atoms are at the centres of the tetrahedra and are not shown. Small open circles ○ Mg atoms at x = zero; small solid circles ● Mg atoms at $x = \frac{1}{2}$ (after Bragg and Brown, 1926).

arranged in approximate hexagonal close packing. In accordance with full orthorhombic symmetry the silicon–oxygen tetrahedra point alternately either way along both the x and y directions. The magnesium atoms do not occupy one set of equivalent lattice positions: half are located at centres of symmetry and half on reflection planes (Fig. 1), the former having as nearest neighbours two oxygens from two adjacent tetrahedra, the latter, two oxygens from one adjacent tetrahedron. In the olivine series the cell dimensions increase with the increasing replacement of the smaller magnesium atom by the larger iron atom. X-ray powder diffraction data for synthetic forsterite and synthetic faya-lite have been determined by Yoder and Sahama (1957) and may be used to estimate the composition of the (Mg,Fe)-olivines. The equation of the deter-minative curve is:

$$\text{Fo (Mol. per cent.)} = 4233 \cdot 91 - 1494 \cdot 59 d_{130}$$

and the fictive end points are forsterite $d_{130} = 2 \cdot 7659$ and fayalite $d_{130} = 2 \cdot 8328$. Depending on the olivine composition the determination is subject to an error of 3 to 4 mol. per cent. (Fig. 2).[1] Tables showing the variation in d values, cell dimensions and density across the olivine series are given by Eliseev (1957). Some cell determinations of members of the olivine group are listed in Table 1.
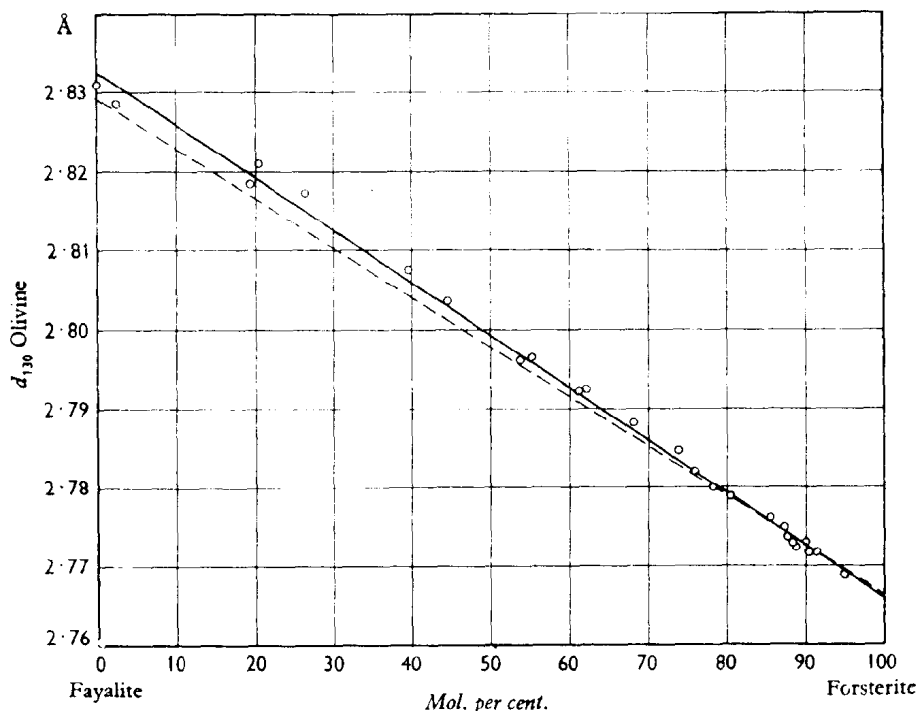
Fig. 2. Olivine X-ray determinative curve (solid line). Circles represent average d_{130} values of analysed specimens used in constructing the curve. Dashed line is assumed linear variation of d_{130} for synthetic olivines using experimentally determined end points (after Yoder and Sahama, 1957).

[1] An alternative curve for determining Mg-rich olivines to $\pm 0 \cdot 4$ mol. per cent., using 2θ 062, has been published (Jackson, E. D., 1960, *U.S. Geol. Surv. Prof. Paper* 400B, p. 432).

Table 1.   CELL PARAMETERS OF OLIVINES

	a (Å)	b (Å)	c (Å)	
Fo_{100}	4·76	10·20	5·99	Swanson & Tatge, 1953.
Fo_{100}	4·756	10·225	5·982	Eliseev, 1957.
Fo_{100}	4·756	10·195	5·981	Yoder & Sahama, 1957.
Fa_{96}	4·812	10·565	6·085	Eliseev, 1957.
Fa_{100}	4·817	10·477	6·105	Yoder & Sahama, 1957.

CHEMISTRY

The composition of olivines varies from forsterite, Mg_2SiO_4, to fayalite, Fe_2SiO_4, there being complete diadochy between Mg and Fe^{+2} (Table 2). In most natural crystals, and particularly in the iron-rich members of the series, there is a little replacement of (Mg,Fe^{+2}) by Mn and Ca.   At the magnesium-rich end of the series Cr and Ni, generally in small amounts, are usually present. For ten olivines examined by Egorova (1938), in which the MgO content varies between 46·72 and 55·74 wt. per cent., the range of NiO is 0·00 to 0·41.   Grigoriev (1937) synthesized a nickel olivine by fusing NiO and SiO_2 with 3 per cent. CaF_2, and the small substitution of (Mg,Fe^{+2}) by Ni in natural olivine is controlled by the small amounts of this ion in the magmas and rocks in which the olivines crystallized.   Ferric iron is reported in many olivine analyses.   The investigation of the $FeO-SiO_2$ system (Bowen and Schairer, 1932) showed that at compositions near Fe_2SiO_4 2·25 per cent. Fe_2O_3 was present in the liquid, and in some olivines small amounts of Fe^{+3} may be primary.   The not infrequent dendritic minute octahedra and opaque plates and wedges oriented
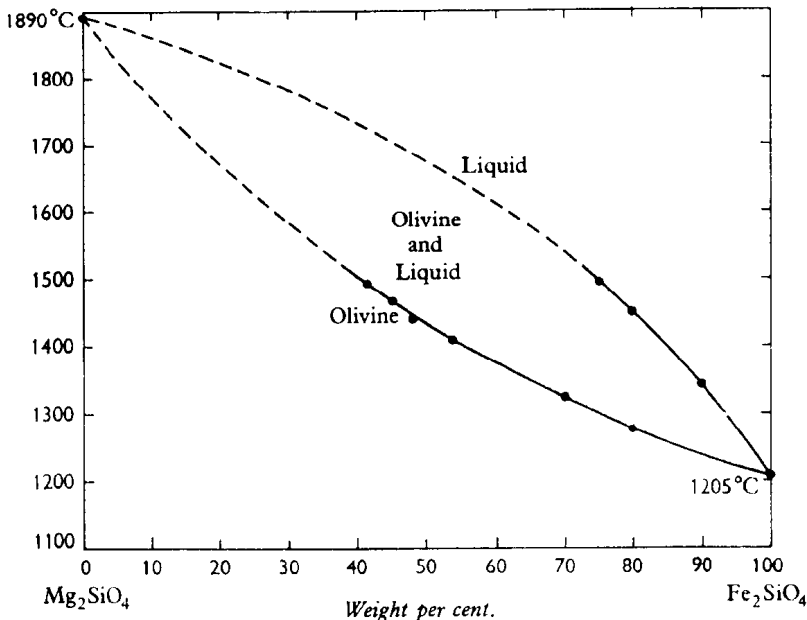

FIG. 3.   Equilibrium diagram of the system Mg_2SiO_4–Fe_2SiO_4 (after Bowen and Schairer, 1935).

parallel to the (001) or (100) planes of the olivine host in some examples have been shown to be magnetite, chromium-rich magnetite, or chromite, and these growths are probably due to the exsolution on cooling of very small amounts of ferric and chromic oxides.   In other olivines, and in all those with relatively high values of Fe^{+3}, the ferric iron is an oxidation product formed by alteration of the olivine.   The alumina and alkalies reported in some olivine analyses are due most probably to impurities in the analysed samples.

The synthetic olivines form a complete solid solution series, the pure magnesium mineral melting at 1890°C. and the pure iron olivine at 1205°C. (Fig. 3). The thermal relationship between forsterite and fayalite illustrates the effect on melting temperatures in a diadochic series of replacing an ion of smaller by one of larger radius.   The cation–oxygen bonds are weaker for the larger cation of the same charge, and as more of the larger cations enter the structure there is a progressive reduction in the melting points of intermediate compositions. Thus the first olivines to separate from a liquid of given composition are richer in Mg than those of later crystallization, and in consequence the larger Fe^{+2}
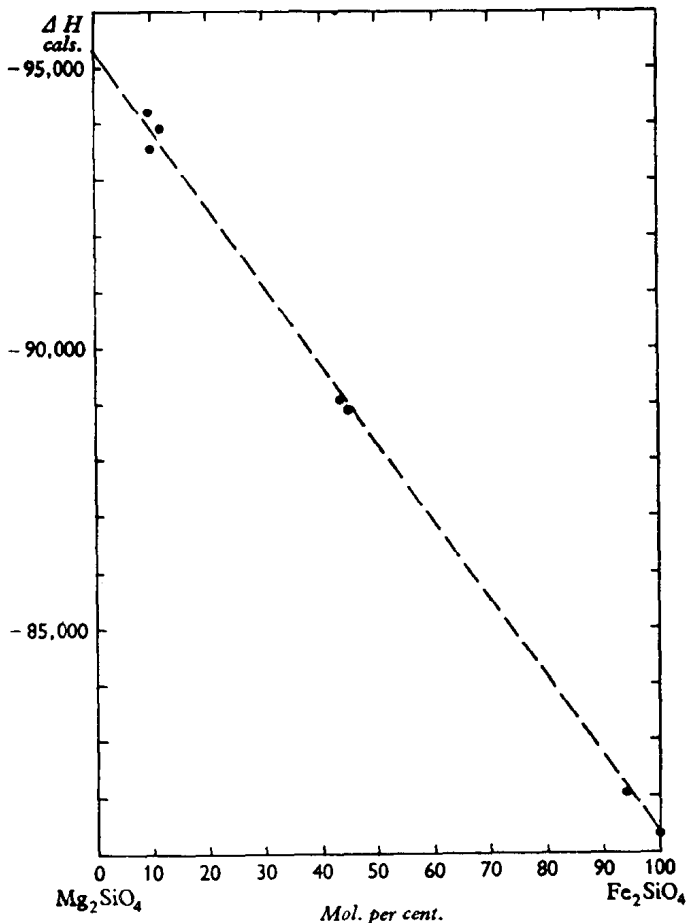

FIG. 4.   Heats of solution in hydrofluoric acid of the (Mg,Fe)-
olivines (after Sahama and Torgeson, 1949).

ions are concentrated in the residual liquids. The perfection of the diadochic replacement of magnesium by ferrous iron in the olivine structure has been substantiated by the measurement of the heats of solution, by solution calorimetry, of a number of olivines varying in molar composition from $Fo_{90.6}Fa_{9.4}$ to $Fo_{0.0}Fa_{100.0}$ (Sahama and Torgeson, 1949). The heats of solution range from the extrapolated value for forsterite of $-95,380$ cal./mole to $-81,330$ cal./mole for synthetic fayalite, and are a linear function of the molar composition (Fig. 4). Thus during the replacement $Mg \rightleftharpoons Fe^{+2}$ perfect thermal equilibrium is maintained, and the heat of isomorphous mixing is zero. The heat of solution of intermediate members of the series may be obtained from the equation:

$$\Delta H \text{ (cal./mol.)} = -81,330 - 140.5 X_{Mg}$$

where X_{Mg} is the mole per cent. Mg_2SiO_4.

Mg-olivines. Magnesium olivines have been synthesized frequently during the investigation of a large number of artificial systems. Many of these studies are of considerable interest and importance in the interpretation of the phase
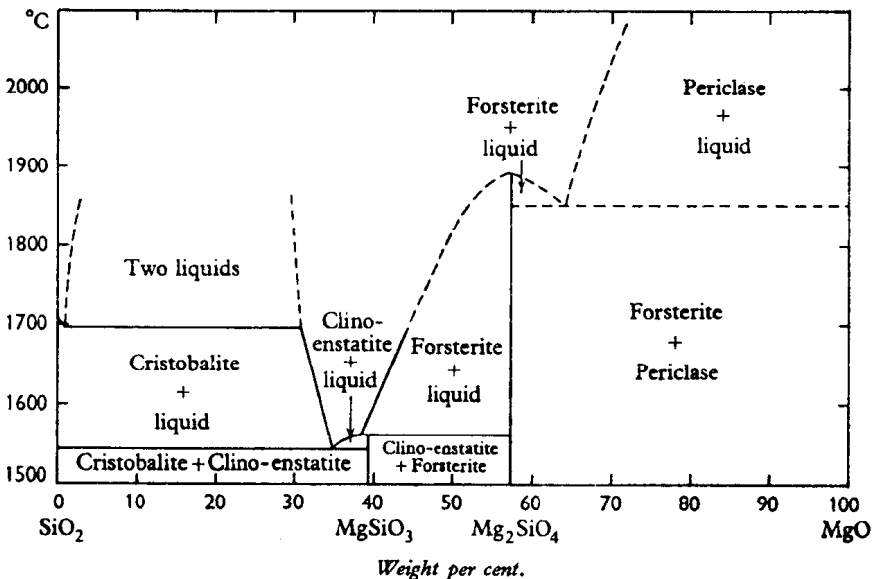
FIG. 5. Equilibrium diagram of the system MgO–SiO_2 (after Bowen and Schairer 1935).

relationships of olivine in natural rocks. In the system MgO–SiO_2 (Fig. 5) the metasilicate $MgSiO_3$, as well as forsterite and SiO_2, crystallizes and forsterite shows an incongruent melting reaction:

$$Mg_2SiO_4 + SiO_2 \rightleftharpoons 2MgSiO_3$$
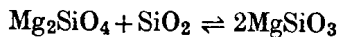
The reaction between olivine crystals and liquid in the system forsterite–silica also occurs in the two ternary systems anorthite–forsterite–silica (Andersen, 1915) and diopside–forsterite–silica (Bowen, 1914) and is confirmed by many natural occurrences. The discontinuous reaction relationship of forsteritic olivine to the magma from which it precipitated is recorded in resorbed crystals and in the reaction rims of peripheral orthopyroxene. Magnetite is sometimes

across the bridge R_3 to some convenient value, such as 1.00 or 3.00 volts, which appears on the voltmeter V. This should have a 0–3-volt range; though even a cheap 3-inch meter suffices for most work, a meter such as the Triplett Model 725 (The Triplett Electrical Instrument Co., Bluffton, Ohio), which has a 6.4-inch mirror scale, is much more accurate and easy to read. If the polarograph is to be used for the measurement of half-wave
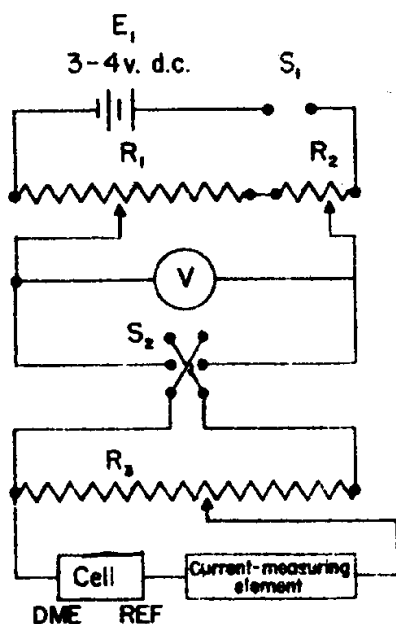
Figure 2. Schematic diagram of manual polarograph. S_1: SPST toggle switch; S_2: DPDT toggle switch; R_1: 500-ohm linear taper potentiometer; R_2: 10-ohm linear taper potentiometer; R_3: voltage divider (see text); V: 0–3 v. d.c. voltmeter.

potentials, the voltmeter must be checked against a precision potentiometer; if very accurate half-wave potential data are required, the mean potential drop across the cell at each point on the c.-v. curve must be measured directly with a precision potentiometer.

The bridge R_3 is used to apply a known fraction of the voltage across it to the portion of the circuit containing the cell and the current-measuring element. So that R_3 will function as an accurate linear voltage divider, its resistance must be very much

when the composition of the liquid is 43·5 wt. per cent. anorthite, 49 per cent. diopside and 7·5 per cent. forsterite (Fig. 7). Spinel crystallizes from some compositions in this system but is a transient phase only, and under equilibrium conditions with falling temperature it is dissolved.

Forsterite has been synthesized from stoichiometric mixtures of its oxides in the presence of water vapour at temperatures of 500°C. and above, at pressures between 2000 and 40,000 lb./in.^{2}. Forsterite is a prominent phase in the ternary system $MgO-SiO_2-H_2O$ (Bowen and Tuttle, 1949). The lowest temperature curve of the *P-T* diagram of this system is of particular paragenetic interest in showing that brucite, serpentine, forsterite and vapour are in equilibrium at temperatures of approximately 400°C. Just above 400°C. at low pressure, and
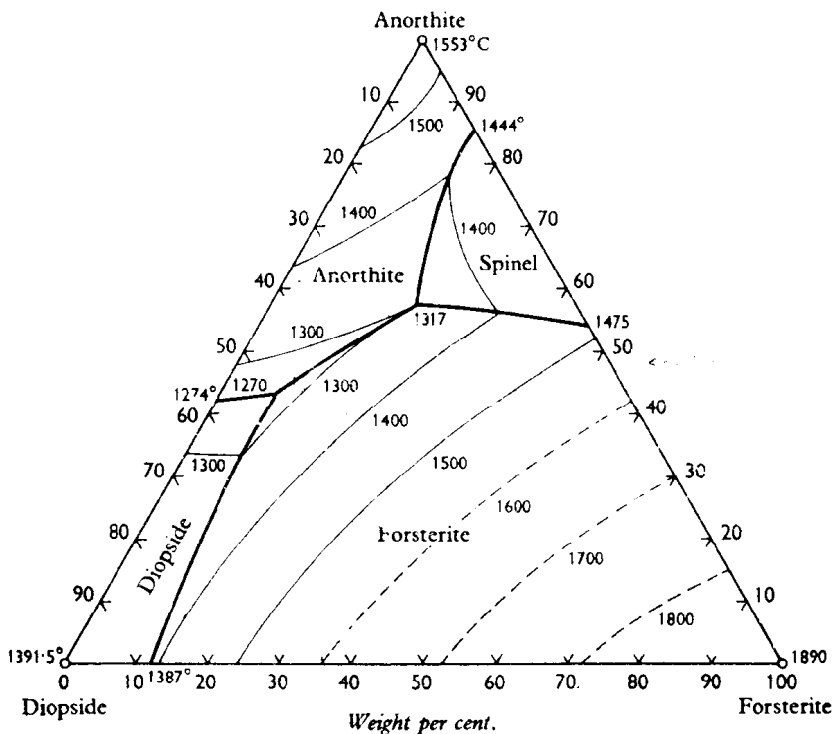
FIG. 7. Equilibrium diagram of the system diopside–forsterite–anorthite. Boundary curves in heavy lines, isotherms in light lines (after Osborn and Tait, 1952).

somewhat higher than 400°C. at higher pressures, forsterite is stable in the presence of water vapour. That the reaction:

$$2Mg(OH)_2 + Mg_6Si_4O_{10}(OH)_8 \rightleftharpoons 4Mg_2SiO_4 + 6H_2O$$

is one of reversible equilibrium has been demonstrated by the serpentinization (formation of brucite and serpentine) of a forsterite crystallized from an anhydrous melt at high temperatures. The replacement of Mg by Fe^{+2} reduces the temperature at which serpentinization takes place, thus in an olivine with 10 mol. per cent. Fe_2SiO_4 the reaction took place at 340°C. (water vapour pressure = 15,000 lb./in.^{2}), while an olivine with 52 mol. per cent. Fe_2SiO_4 was not serpentinized at 295°C. and 15,000 lb./in.^{2} water vapour pressure.

Investigations of the quaternary system $MgO-Al_2O_3-SiO_2-H_2O$ (Yoder, 1952; Roy and Roy, 1955) have confirmed the main results of the above ternary system, forsterite crystallizing from powdered mixes and glasses of appropriate composition at approximately 430°C. and under excess water vapour pressure. In this system two sets of equilibrium associations have been recognized, (a) phase associations in equilibrium in the presence of excess water, (b) associations in equilibrium under "water deficient" conditions, *i.e.* those in which there is insufficient water to convert all the phases to hydrates. Forsterite in equilib-

Table 2.   OLIVINE ANALYSES

	1.	2.	3.	4.	5.
SiO_2	41·72	41·07	40·70	41·29	40·96
TiO_2	—	0·05	tr.	tr.	0·01
Al_2O_3	—	0·56	0·60	0·27	0·21
Fe_2O_3	—	0·65	1·60	0·22	0·00
FeO	1·11	3·78	4·47	7·09	7·86
MnO	—	0·23	0·32	0·42	0·13
MgO	57·83	54·06	51·84	51·07	50·45
CaO	—	0·00	tr.	0·13	0·15
Na_2O	—	—	tr.	—	0·01
K_2O	—	—	tr.	—	0·00
H_2O^+	—	0·05	⎰0·51	—	0·29
H_2O^-	—	0·00	⎱	—	—
Total	100·66	100·45	100·04	100·49	100·35
α	—	—	1·649		—
β	—	—	—	1·664	—
γ	—	—	1·689	1·685	—
$2V_\gamma$	86° 20′	—	86°	87°	—
D	3·22	—	3·187	3·22	3·313

NUMBERS OF IONS ON THE BASIS OF 4 OXYGENS

	1.	2.	3.	4.	5.
Si	0·978	0·979	0·983	0·996	0·996
Al	—	0·016	0·017	0·007	—
Ti	—	0·001	—	—	—
Fe^{+3}	—	0·012	0·030	0·004	—
Mg	2·021 ⎱2·04	1·920 ⎱2·03	1·867 ⎱2·01	1·836 ⎱2·00	1·828 ⎱2·0
Fe^{+2}	0·022	0·075	0·090	0·143	0·160
Mn	—	0·005	0·007	0·009	0·003
Ca	—	—	—	0·003	0·004
Atomic ratios					
Mg	98·9	96·2	95·4	92·8	92·0
Fe^{+2}	1·1	3·8	4·6	7·2	8·0

1. Forsterite, crystalline limestone, Mogok Mining District, Burma (Adams and Graham, 1926).
2. Olivine, with clinohumite in limestone, Ojama, Finland (Sahama, 1953).
3. Olivine, forsterite–mica–magnetite rock, Vostochyni mine, Ensky, U.S.S.R. (Sobolev, 1947).
4. Olivine, tuff, Alban Hills, Italy (Fornaseri, 1951).
5. Olivine, dunite, Dun Mountain, New Zealand (Ross *et al.*, 1954).   Anal. M. D. Foster (Includes Cr_2O_3 0·02, CoO 0·01, NiO 0·25).

* Includes Ni 0·005.

rium with cordierite and spinel can co-exist at water vapour pressures above 15,000 lb./in.² and a temperature of 680°C.

$$6Mg_5Al_2Si_3O_{10}(OH)_8 \rightleftharpoons 12Mg_2SiO_4 + 4MgAl_2O_4 + Mg_2Al_4Si_5O_{18} + SiO_2 + 24H_2O$$

$$\underset{\text{clinochlore}}{} \qquad \underset{\text{forsterite}}{} \quad \underset{\text{spinel}}{} \qquad \underset{\text{cordierite}}{}$$
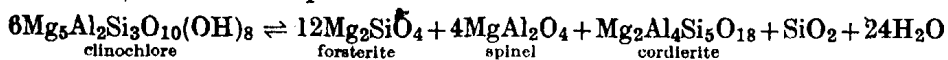
Forsterite has also been crystallized from gels precipitated from solutions of Na_2SiO_3 and $MgCl_2$ (Sabatier, 1950).

Table 2 contd.   OLIVINE ANALYSES

	6.	7.	8.	9.	10.
SiO_2	40·84	40·30	39·12	40·21	39·87
TiO_2	0·04	0·15	0·09	0·00	0·03
Al_2O_3	0·19	0·25	0·58	—	0·00
Fe_2O_3	0·13	0·00	1·69	—	0·86
FeO	8·18	10·26	10·76	12·57	13·20
MnO	0·17	0·09	0·16	—	0·22
MgO	50·27	48·60	46·51	47·49	45·38
CaO	—	0·07	0·48	0·20	0·25
Na_2O	—	0·04	0·15	—	0·04
K_2O	—	0·03	0·06	—	0·01
H_2O+	0·37	}0·33	0·21	—	0·33
H_2O-	—		0·07	—	0·10
Total	100·38	100·56	99·95	100·47	100·30
α	1·649	—	1·656	—	1·6626
β	1·666	—	1·674	—	—
γ	1·684	—	1·695	—	1·6990
$2V_\gamma$	—	—	89°	—	90°
D	3·30	3·349	3·33	3·38	3·44

NUMBERS OF IONS ON THE BASIS OF 4 OXYGENS

	6.	7.	8.	9.	10.
Si	0·994	0·990	0·975	0·993	0·997
Al	0·005⎫	0·007⎫	0·016⎫	—⎫	—⎫
Ti	0·001⎪	0·003⎪	0·002⎪	—⎪	0·001⎪
Fe^{+3}	0·002⎪	—⎪	0·031⎪	—⎪	0·016⎪
Mg	1·824 ⎬2·02†	1·779 ⎬2·02‡	1·728 ⎬2·03§	1·748 ⎬2·01	1·692 ⎬2·00‖
Fe^{+2}	0·177⎪	0·211⎪	0·224⎪	0·260⎪	0·276⎪
Mn	0·004⎪	0·002⎪	0·003⎪	—⎪	0·005⎪
Ca	—⎭	0·002⎭	0·013⎭	0·005⎭	0·007⎭
Atomic ratios					
Mg	91·1	89·4	88·5	87·1	86·0
Fe^{+2}	8·9	10·6	11·5	12·9	14·0

6. Olivine, dunite, Little Castle Creek, California (Hawkes, 1946).   Anal. M. Fleischer (Includes Ni 0·19).

7. Olivine, nodules in basalt, Ichinomegata, Japan (Ross *et al.*, 1954).   Anal. M. D. Foster (Includes NiO 0·41, Cr_2O_3 0·03).

8. Olivine, melilite basalt, Spiegel river, Cape Province (Mathias, 1949).   Anal. Geochemical Laboratories   (Includes P_2O_5 0·07).

9. Olivine, pallasite meteorite, Huckitta, central Australia (Madigan, 1939).   Anal. A. R. Alderman.

10. Olivine, allivalite, Rhum, Scotland (Brown, 1956).   Anal. G. M. Brown   (Includes P_2O_5 0·01, Cr_2O_3 trace).

† Includes Ni 0·004.

‡ Includes Na 0·002, K 0·001, Ni 0·008, Cr 0·001.

§ Includes Na 0·007, K 0·002.

‖ Includes Na 0·002.

Table 2 contd.   OLIVINE ANALYSES

	11.	12.	13.	14.	15.
SiO_2	39·31	37·33	38·11	34·96	34·04
TiO_2	0·06	0·09	tr.	0·00	0·43
Al_2O_3	1·68	0·18	0·00	0·00	0·91
Fe_2O_3	----	1·60	0·15	0·00	1·46
FeO	19·84	21·58	31·48	36·77	40·37
MnO	0·17	0·27	0·22	0·52	0·68
MgO	37·74	38·13	30·50	27·04	20·32
CaO	0·87	0·38	0·02	0·00	0·81
Na_2O	---	0·03	----	----	. ---
K_2O	--	0·05	---	- ..	-- .
H_2O	---	0·17	----	----	0·09
H_2O^-	----	0·04	-	----	----
Total	99·67	99·94	100·48	"100·58"	99·11
α	1·675	1·683	1·710	1·721	1·742
β	1·694	1·704	1·733	1·750	.
γ	1·712	1·722	1·748	1·765	---
$2V_\gamma$	90°	95°	101°	95°	102°–104°
D	. . .	3·53	3·69	. . .	----

NUMBERS OF IONS ON THE BASIS OF 4 OXYGENS

Si	1·010	0·979	1·026	0·989	0·990
Al	0·051 ⎫	0·005 ⎫	— ⎫	.... ⎫	0·032 ⎫
Ti	0·001 ⎪	0·002 ⎪	---- ⎪	— ⎪	0·009 ⎪
Fe^{+3}	— ⎪	0·031 ⎪	0·003 ⎪	.... ⎪	0·032 ⎪
Mg	1·446 ⎬1·95	1·491 ⎬2·02¶	1·226 ⎬1·95	1·140 ⎬2·02	0·881 ⎬1·98
Fe^{+2}	0·427 ⎪	0·474 ⎪	0·709 ⎪	0·870 ⎪	0·983 ⎪
Mn	0·004 ⎪	0·006 ⎪	0·005 ⎪	0·012 ⎪	0·017 ⎪
Ca	0·024 ⎭	0·011 ⎭	0·006 ⎭	— ⎭	0·025 ⎭

Atomic ratios					
Mg	77·2	75·9	63·3	59·7	47·3
Fe^{+2}	22·8	24·1	36·7	40·3	52·7

11. Olivine, basalt, Buffalo Buttes, Colorado (Larsen *et al.*, 1936).   Anal. F. A. Gonyer.
12. Olivine, melilite basalt, Klaasvoogds, Cape Province (Mathias, 1949).   Anal. Geochemical
    Laboratories   (Includes P_2O_5 0·09).
13. Olivine, hypersthene–olivine gabbro, Kangerdlugssuaq, east Greenland (Deer and Wager,
    1939).   Anal. W. A. Deer.
14. Hyalosiderite, gabbro, Black Cascade, Tripyramid Mountain, New Hampshire (Chapman and
    Williams, 1935).   Anal. F. A. Gonyer.
15. Olivine, olivine gabbro, Camas Mòr, Muck. Scotland (Tilley, 1952).   Anal. J. H. Scoon.

¶ Includes Na 0·001, K 0·001.

Table 2 contd. OLIVINE ANALYSES

	16.	17.	18.	19.	20.	21.
SiO_2	31·85	30·42	30·15	29·83	29·75	29·87
TiO_2	0·01	1·20	0·20	—	—	0·08
Al_2O_3	tr.	0·50	0·07	—	0·00	—
Fe_2O_3	0·11	0·00	0·43	—	0·83	12·07
FeO	58·64	57·62	65·02	69·48	66·10	54·88
MnO	0·85	—	1·01	0·28	3·20	0·02
MgO	8·49	8·17	1·05	—	—	2·54
CaO	0·18	1·32	2·18	—	—	0·20
Na_2O	—	—	—	—	—	0·71
K_2O	—	—	—	—	—	0·08
H_2O^+	—	0·18	—	0·34	0·19	—
H_2O^-	—	0·21	—	0·04	—	—
Total	100·13	99·62	100·11	99·97	100·07	100·45
α	1·788	1·787	1·827	1·8235	—	1·870
β	1·828	1·821	1·869	1·8649	—	—
γ	1·840	1·838	1·879	1·8770	—	1·908
$2V_\gamma$	122°	—	132°	130°	—	131°
D	4·15	—	—	4·215	—	4·1

NUMBERS OF IONS ON THE BASIS OF 4 OXYGENS

	16.	17.	18.	19.	20.	21.
Si	1·007	0·971	1·002	1·011	1·010	0·962
Al	—	0·019	0·003	—	—	—
Ti		0·029	0·005			0·002
Fe^{+3}	0·002	—	0·011	—	0·020	0·292
Mg	0·400 }1·98	0·389 }2·02	0·052 }1·99	— }1·98	— }1·99	0·122 }1·95**
Fe^{+2}	1·552	1·539	1·808	1·970	1·877	1·479
Mn	0·023	—	0·028	0·008	0·092	0·001
Ca	0·006	0·045	0·078	—	—	0·007

Atomic ratios

	16.	17.	18.	19.	20.	21.
Mg	20·5	20·2	2·8	0·0	0·0	—
Fe^{+2}	79·5	79·8	97·2	100·0	100·0	—

16. Olivine, ferrohortonolite ferrogabbro, Kangerdlugssuaq, east Greenland (Deer and Wager 1939). Anal. W. A. Deer.

17. Olivine, iron-rich diabase, Beaver Bay, Minnesota (Muir, 1954). Anal. I. D. Muir.

18. Olivine, fayalite ferrogabbro, Kangerdlugssuaq, east Greenland (Deer and Wager, 1939). Anal. W. A. Deer.

19. Fayalite, metamorphosed iron-rich sediments, Brocken, Harz, Germany (Ramdohr, 1927). Anal. J. Jakob.

20. Fayalite, Rockport, Massachusetts (Bowen *et al.*, 1933a). Anal. E. Posnjak.

21. Talasskite, pegmatite, Talassa Valley, Siberia (Nikitin, 1936).

** Includes Na 0·045, K 0·002.