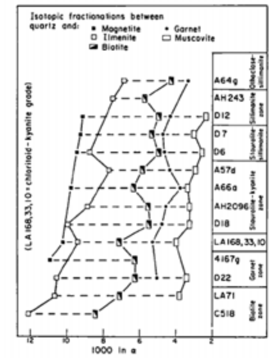

Fig. 18. Isotopic fractionations in non-retrograded rocks correlated with metamorphic grade.

From Garlick and Epstein 1967

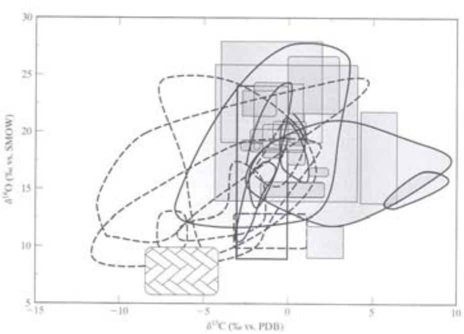

FIGURE 12.1: Compilation of literature on marbles. There is a clustering of $\delta^{18}O$ and $\delta^{13}C$ values around 20‰ (SMOW) and 0‰ (PDB), respectively. Siliceous marbles and calc–silicates (shown as dashed lines) have lower $\delta^{18}O$ and, especially, $\delta^{13}C$ values. Also shown is the igneous carbonatite box (hatched area). (Each shaded area represents the range of values from an individual publication).

Quartz
Feldspar
Muscovite
Kyanite
Hornblende
Garnet
Biotite
Chlorite
Ilmenite
Magnetite

Increasing tendency to concentrate O^{18}

From Garlick and Epstein 1967

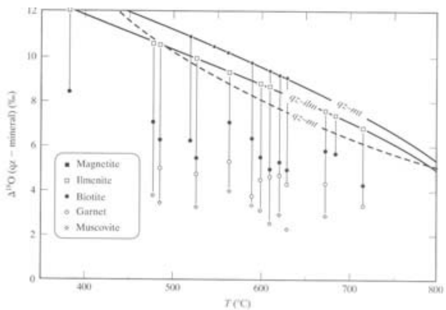

FIGURE 12.13: Garlick and Epstein's (1967) measured fractionations of minerals relative to quartz for several metamorphic terranes. The authors noted that there is a general decrease in fractionation as temperatures increase. The data are plotted relative to temperature, as determined by the quartz–magnetite fractionation available at the time (top curve). A more recent calibration is shown as the dashed line. In general, fractionations for all minerals decrease with increasing grade and there is an isotopic enrichment of one mineral relative to another. Note that the smooth variation in magnetite is a result of the construction of the curve, where the X-axis is defined by the $\Delta^{18}O_{qz-mt}$ value.

TABLE 12.P1: Some commonly used coefficients of oxygen isotope fractionation relative to quartz. In form of equation 12.11 (after Javoy, 1977).

Mineral	a	b
Plagioclase	0.97 + 1.04β	0
Olivine	2.75	0
Muscovite	2.2	−0.6
Biotite	3.69	−0.6
Amphibole	3.15	−0.3
Ilmenite	5.29	0
Magnetite	5.57	0
Garnet	2.88	0
Chlorite	5.44	−1.63

β is the fraction of anorthite in plagioclase.

Isotherm

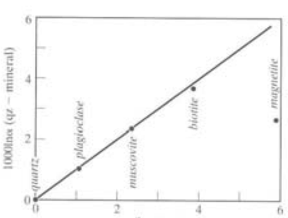

FIGURE 12.14: Typical isotherm plot for coexisting minerals. All mineral data are plotted relative to quartz (although the choice of reference mineral is arbitrary and will not change the result). In the example illustrated, all minerals except magnetite define an isotherm, and magnetite is clearly out of equilibrium with the other phases. The slope of the best-fit line is equal to $10^6/T^2$ (with T in kelvins).

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Oxygen isotope ratios in coexisting minerals of regionally metamorphosed rocks

G. DONALD GARLICK* and SAMUEL EPSTEIN†

Increasing	↑	Garnet
"isotopic	|	Quartz
stability"	|	Ilmenite, muscovite, biotite

Geospeedometry

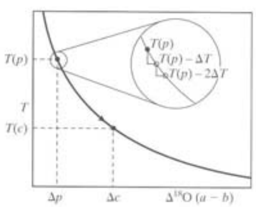

FIGURE 12.16: Schematic plot of $\Delta^{18}O$ value for phases a and b vs. temperature. The initial, peak temperature is $T(p)$. If the sample is cooled rapidly to room temperature, it will preserve fractionations Δp, corresponding to equilibrium at $T(p)$. If the sample is cooled slowly, it will continue to reequilibrate during cooling, following the equilibrium fractionation-vs.-temperature path (e.g., $T(p)$ to $T(p) - \Delta T$, etc.). The sample will continue to reequilibrate until the closure temperature $T(c)$ is reached, at which point intracrystalline diffusion effectively ceases.

[8] Diffusional exchange in this case is limited by the **intractrystalline** diffusion rate, or self-diffusion rate. This is the diffusion rate of oxygen within a crystal. The **intercrystalline**, or grain-boundary, diffusion rate, which allows oxygen to migrate between phases, is considered to be orders of magnitude more rapid than the intracrystalline diffusion rate. Limited experimental studies support this assumption (e.g., Farver and Yund, 1991a).