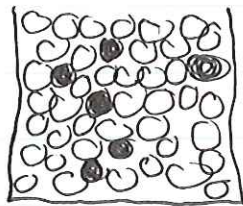


K-Ar and Rb-Sr geochronology — the age of the Earth :

¹⁴C cannot be used to date rocks and meteorites — with few exceptions older than 40,000 years.

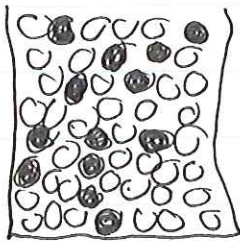
Basis of all radiometric dating:



$t = 0$

- P_0 parent atoms
- D_0 daughter atoms

Later, at time t :



$t = t$

$$P_t = P_0 e^{-\lambda t}$$

$$D_t = D_0 + P_0 (1 - e^{-\lambda t})$$

$$t = \frac{1}{\lambda} \ln \left(\frac{P_0}{P_t} \right)$$

write on board before lecture

Need to know P_0 — e.g. ¹⁴C — $P_0 = (^{14}\text{C})_{\text{atmosphere}}$

$$t = \frac{1}{\lambda} \ln \left(\frac{D_t - D_0}{P_t} + 1 \right)$$

Can measure D_t, P_t now — but must also know D_0 or figure out how to avoid needing to know it.

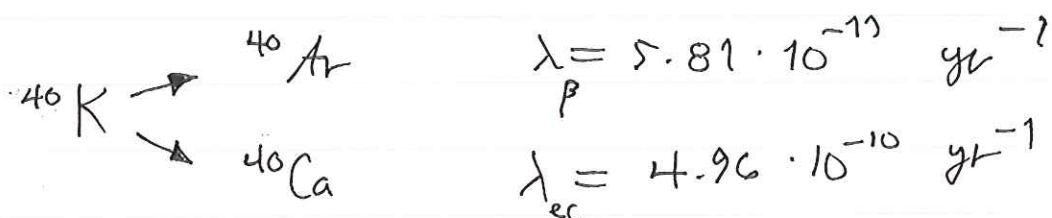
should say... or be able to determine it

Table 3.1 lists common decay schemes used to date rocks & meteorites

Discuss ~~the~~ first two in greater detail

K-Ar dating

A complication — has two decay branches



Respective half lives

$$\tau_{1/2} ({}^{40}\text{Ar}) = \text{~~11.9 b.y.~~} 11.9 \text{ b.y.}$$

$$\tau_{1/2} ({}^{40}\text{Ca}) = 1.4 \text{ b.y.}$$

$$\tau_{1/2}^{\text{total}} = \frac{.693}{\lambda_{\text{ec}} + \lambda_{\beta}} = 1.25 \text{ b.y.}$$

Useful for rocks 1 m.y. — age of \oplus
 Age equation must be modified
 to account for two branches

$$t = \frac{1}{\lambda_{\text{ec}} + \lambda_{\beta}} \ln \left(\frac{{}^{40}\text{Ar}}{{}^{40}\text{K}} \frac{\lambda_{\text{ec}} + \lambda_{\beta}}{\lambda_{\text{ec}}} + 1 \right)$$

Go to
 page 3
 first

$$t = 1.804 \cdot 10^9 \ln \left(9.54 \frac{^{40}\text{Ar}}{^{40}\text{K}} + 1 \right) \text{ yrs}$$



About 9/10 of every ^{40}K \rightarrow ~~^{40}Ca~~ ^{40}Ca ,
 only 1/10 \rightarrow ^{40}Ar ; hence, must
 multiply ^{40}Ar by 9.54

Before giving age equation

\rightarrow K-Ar dating used to date
crystallization ages of igneous rocks.
 when did they crystallize from melt.

^{40}K common element in many minerals,
 e.g. K-feldspar, KAlSi_3O_8

Ca is even more abundant and ^{40}Ca is
 97% of all Ca — so it is very
 difficult to find D_0 for the ^{40}Ca
 decay branch.

But Ar is a noble gas — see periodic
 table — volatile — escapes upon melting.

Hence $D_0 = 0$ at time of crystallization —
 noble gas is not accommodated into
 crystal lattice. Any ^{40}Ar formed by
 decay is, however, trapped in the
 lattice like a bird in a cage.

TABLE 3.1
Principal Parent and Daughter Isotopes Used to
Determine the Ages of Rocks and Minerals

Parent isotope (radioactive)	Daughter isotope (stable)	Half-life (Ma)	Decay constant (yr^{-1})
^{40}K	$^{40}\text{Ar}^a$	1,250	5.81×10^{-11}
^{87}Rb	^{87}Sr	48,800	1.42×10^{-11}
^{147}Sm	^{143}Nd	106,000	6.54×10^{-12}
^{176}Lu	^{176}Hf	35,900	1.93×10^{-11}
^{187}Re	^{187}Os	43,000	1.612×10^{-11}
^{232}Th	^{208}Pb	14,000	4.948×10^{-11}
^{235}U	^{207}Pb	704	9.8485×10^{-10}
^{238}U	^{206}Pb	4,470	1.55125×10^{-10}

^a ^{40}K also decays to ^{40}Ca , for which the decay constant is $4.962 \times 10^{-10} \text{ yr}^{-1}$, but that decay is not used for dating. The half-life is for the parent isotope and so includes both decays.

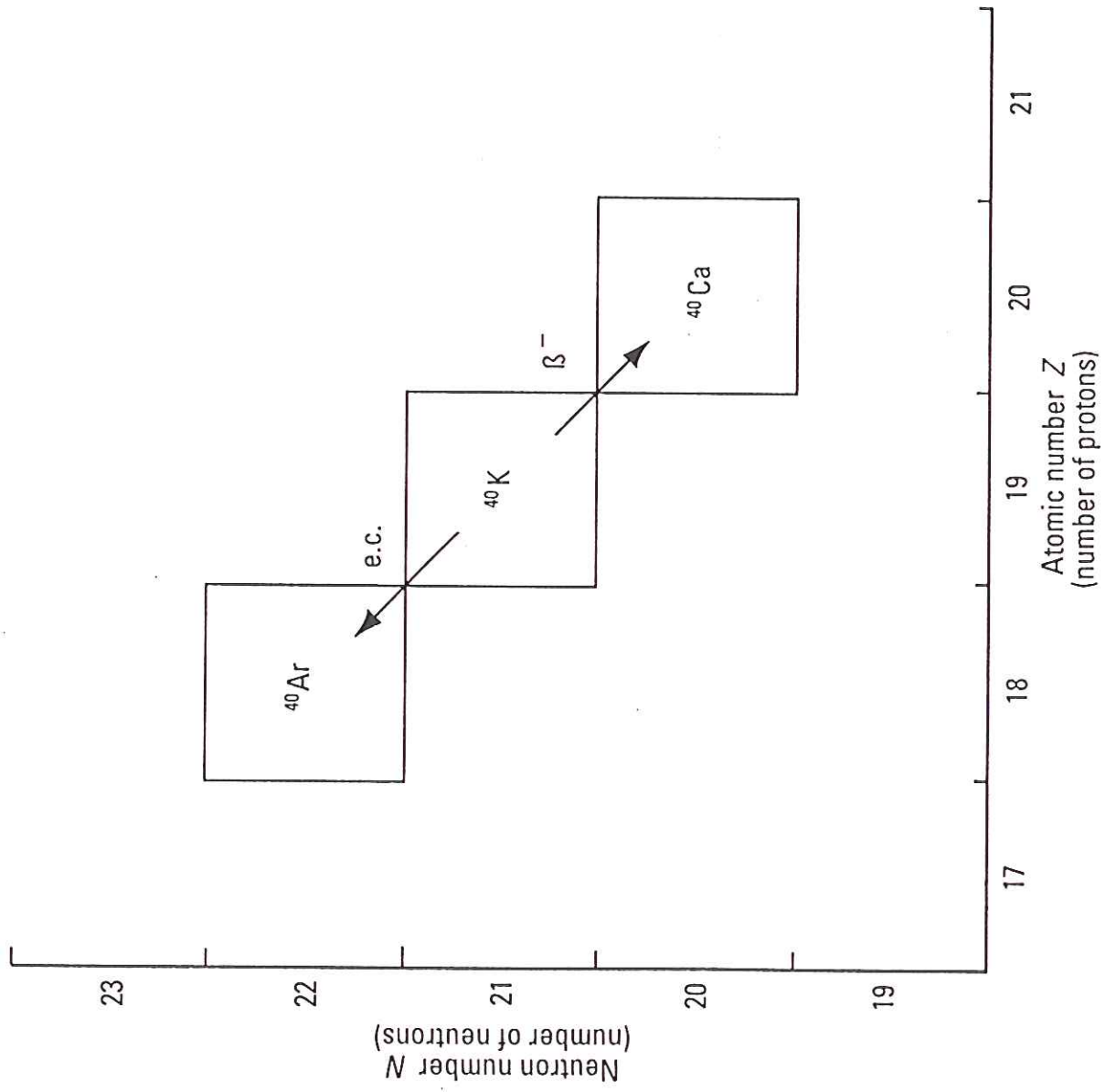


Fig. 3.3. ^{40}K decays by β^- emission to ^{40}Ca and by electron capture to ^{40}Ar in a ratio of 8.5 to 1. Only the decay to Ar is used for radiometric dating.

PERIODIC TABLE OF THE ELEMENTS

Mn — Chemical Symbol
 25 — Atomic Number
 Manganese — Element Name
 54.94 — Atomic Weight

Strong tendency for outermost electrons to be lost to make full outer shell		Transition elements: valence electrons not in outer shell										Noble gases: outer shells filled; no tendency to gain or lose electrons																																																													
H 1 Hydrogen 1.01	Li 3 Lithium 6.94	Be 4 Beryllium 9.01	B 5 Boron 10.81	C 6 Carbon 12.01	N 7 Nitrogen 14.01	O 8 Oxygen 16.00	F 9 Fluorine 19.00	Ne 10 Neon 20.18	Na 11 Sodium 22.99	Mg 12 Magnesium 24.31	Al 13 Aluminum 26.98	Si 14 Silicon 28.09	P 15 Phosphorus 30.97	S 16 Sulfur 32.06	Cl 17 Chlorine 35.45	Ar 18 Argon 39.95	K 19 Potassium 39.10	Ca 20 Calcium 40.08	Sc 21 Scandium 44.96	Ti 22 Titanium 47.90	V 23 Vanadium 50.94	Cr 24 Chromium 52.00	Mn 25 Manganese 54.94	Fe 26 Iron 55.85	Co 27 Cobalt 58.93	Ni 28 Nickel 58.70	Cu 29 Copper 63.55	Zn 30 Zinc 65.38	Rb 37 Rubidium 85.47	Sr 38 Strontium 87.62	Y 39 Yttrium 88.91	Zr 40 Zirconium 91.22	Nb 41 Niobium 92.91	Mo 42 Molybdenum 95.94	Tc 43 Technetium (98)	Ru 44 Ruthenium 101.07	Rh 45 Rhodium 102.91	Pd 46 Palladium 106.4	Ag 47 Silver 107.87	Cd 48 Cadmium 112.41	In 49 Indium 114.82	Sn 50 Tin 118.69	Sb 51 Antimony 121.75	Te 52 Tellurium 127.60	I 53 Iodine 126.90	Xe 54 Xenon 131.30	Cs 55 Cesium 132.91	Ba 56 Barium 137.33	* (see below)	Hf 72 Hafnium 178.49	Ta 73 Tantalum 180.95	W 74 Tungsten 183.85	Re 75 Rhenium 186.21	Os 76 Osmium 190.2	Ir 77 Iridium 192.22	Pt 78 Platinum 195.09	Au 79 Gold 196.97	Hg 80 Mercury 200.59	Tl 81 Thallium 204.37	Pb 82 Lead 207.2	Bi 83 Bismuth 208.98	Po 84 Polonium (209)	At 85 Astatine (210)	Rn 86 Radon (222)	Fr 87 Francium (223)	Ra 88 Radium 226.03	** (see below)	Rf 104 Rutherfordium 261.11	Db 105 Dubnium 262.11	Sg 106 Seaborgium (263)	Bh 107 Bohrium (262)	Hs 108 Hassium (264)	Mt 109 Meitnerium (266)
IA	IIA	IIIB	IVB	VB	VIB	VII B	VIII B	VIII B	IB	IIB	IIIA	IVA	VA	VIA	VIIA	VIIIA																																																									

Lanthanide (Rare Earth) Elements														Actinide Elements															
La *57 Lanthanum 138.91	Ce 58 Cerium 140.12	Pr 59 Praseodymium 140.91	Nd 60 Neodymium 144.24	Pm 61 Promethium (145)	Sm 62 Samarium 150.4	Eu 63 Europium 151.96	Gd 64 Gadolinium 157.25	Tb 65 Terbium 158.93	Dy 66 Dysprosium 162.50	Ho 67 Holmium 164.93	Er 68 Erbium 167.26	Tm 69 Thulium 168.93	Yb 70 Ytterbium 173.04	Lu 71 Lutetium 174.97	Ac **89 Actinium 227.03	Th 90 Thorium 232.04	Pa 91 Protactinium 231.04	U 92 Uranium 238.03	Np 93 Neptunium 237.05	Pu 94 Plutonium (244)	Am 95 Americium (243)	Cm 96 Curium (247)	Bk 97 Berkelium (247)	Cf 98 Californium (251)	Es 99 Einsteinium (252)	Fm 100 Fermium (257)	Md 101 Mendelevium (258)	No 102 Nobelium (259)	Lr 103 Lawrencium (260)

TABLE 3.2
Natural Abundances of the Isotopes Used in Radiometric Dating

Isotope	Abundance (%)	Isotope	Abundance (%)
^{39}K	93.26	^{36}Ar	0.337
^{40}K	0.0117	^{38}Ar	0.063
^{41}K	6.73	^{40}Ar	99.60
^{85}Rb	72.17	^{84}Sr	0.56
^{87}Rb	27.83	^{86}Sr	9.87
		^{87}Sr	7.04
		^{88}Sr	82.53
^{144}Sm	3.0	^{142}Nd	27.3
^{147}Sm	14.9	^{143}Nd	12.3
^{148}Sm	11.2	^{144}Nd	23.8
^{149}Sm	13.8	^{145}Nd	8.3
^{150}Sm	7.4	^{146}Nd	17.1
^{152}Sm	26.8	^{148}Nd	5.7
^{154}Sm	22.9	^{150}Nd	5.6
^{175}Lu	97.4	^{174}Hf	0.17
^{176}Lu	2.6	^{176}Hf	5.2
		^{177}Hf	18.5
		^{178}Hf	27.2
		^{179}Hf	13.8
		^{180}Hf	35.1
^{185}Re	37.40	^{184}Os	0.02
^{187}Re	62.6	^{186}Os	1.6
		^{187}Os	1.6
		^{188}Os	13.3
		^{189}Os	16.1
		^{190}Os	26.4
		^{192}Os	41.0
^{232}Th	100.0	^{204}Pb	1.4
		^{206}Pb	25.2
^{234}U	0.0057	^{207}Pb	21.7
^{235}U	0.72	^{208}Pb	51.7
^{238}U	99.27		

NOTE: Abundances are for the Earth's crust except for argon, which is for the atmosphere. The isotopic abundances for those elements that include a daughter isotope vary because of decay of the corresponding parent isotope. The isotope pairs used in radiometric dating are indicated by arrows.

SOURCES: Lederer, Holland, and Perlman, 1967; Faure, 1986.

K abundances

^{39}K	93%
^{40}K	0.01% only
^{41}K	7%

Ar abundances

^{40}Ar 99.6%

↑ i.e. ~ 1% of atmosphere — all from decay of ^{40}K

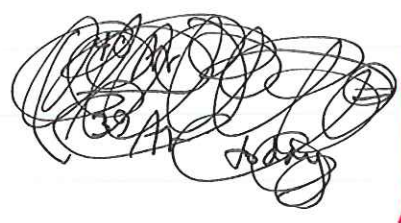
K-Ar method is useful because:

- K abundant in most rocks
- hence plenty of ^{40}K even though least abundant isotope
- argon lost easily at $T \sim T_{\text{melt}}$

write on board

Mid 1960's — an improvement to the method

Why is $\frac{^{40}\text{K}}{^{39}\text{K}}$ constant?
 Because K does not leave when the K-Ar clock is reset.



$$\left(\frac{^{40}\text{K}}{^{39}\text{K}} \right)_{\text{today}} = \frac{.0117}{93.26} = \frac{1}{7970}$$

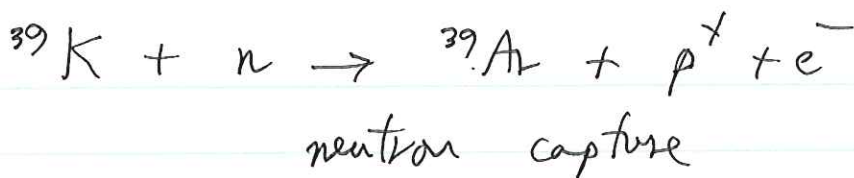
This present-day ratio the same in all rocks — rocks of all ages — so measuring ^{39}K equivalent to measuring ^{40}K

Irradiate sample with fast neutrons at a reactor

since $^{40}\text{K} \rightarrow ^{40}\text{Ca}$
of ^{40}K

$$t = 1.804 \cdot 10^9 \ln \left(9.54 \frac{^{40}\text{Ar}}{^{39}\text{K}} + 1 \right)$$

5



$$^{39}\text{K} = ^{39}\text{Ar} \times \text{conversion efficiency}$$

Thus, ^{39}Ar can provide a surrogate for ^{40}K

Age equation now

$$t = 1.804 \cdot 10^9 \ln \left(J \frac{^{40}\text{Ar}}{^{39}\text{Ar}} + 1 \right)$$

$$\frac{1}{\lambda} = \frac{1}{\lambda_{\text{ec}}} + \frac{1}{\lambda_{\beta}}$$

accounts for both $^{40}\text{K}/^{39}\text{K} = 1/7970$ and conversion efficiency

$$J = \frac{7970}{\text{conv. eff.}}$$

To determine J ~~known~~ irradiate a sample of known age at same time

$$J = \frac{e^{\lambda t_{\text{known}}} - 1}{(^{40}\text{Ar}/^{39}\text{Ar})_{\text{known}}}$$

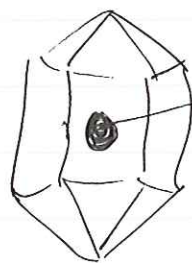
$$t_{\text{unknown}} = \frac{1}{\lambda} \ln \left(J \frac{^{40}\text{Ar}}{^{39}\text{Ar}} + 1 \right)$$

Advantage of method — can see if sample has been disturbed by subsequent reheating.

Show figure with undisturbed & disturbed age spectrum

Plot age t versus % ^{39}Ar released

Sample (c) lost 15% of its argon at time t' — but plateau gives original formation age



last Ar released from center of xtal where it is tightly bound

Advantages :

- $^{40}\text{Ar}/^{39}\text{Ar}$ can be measured in a single mass spectrometer run — just irradiate & step heat
- can do on mineral separates or on whole rock
- can see evidence of reheating

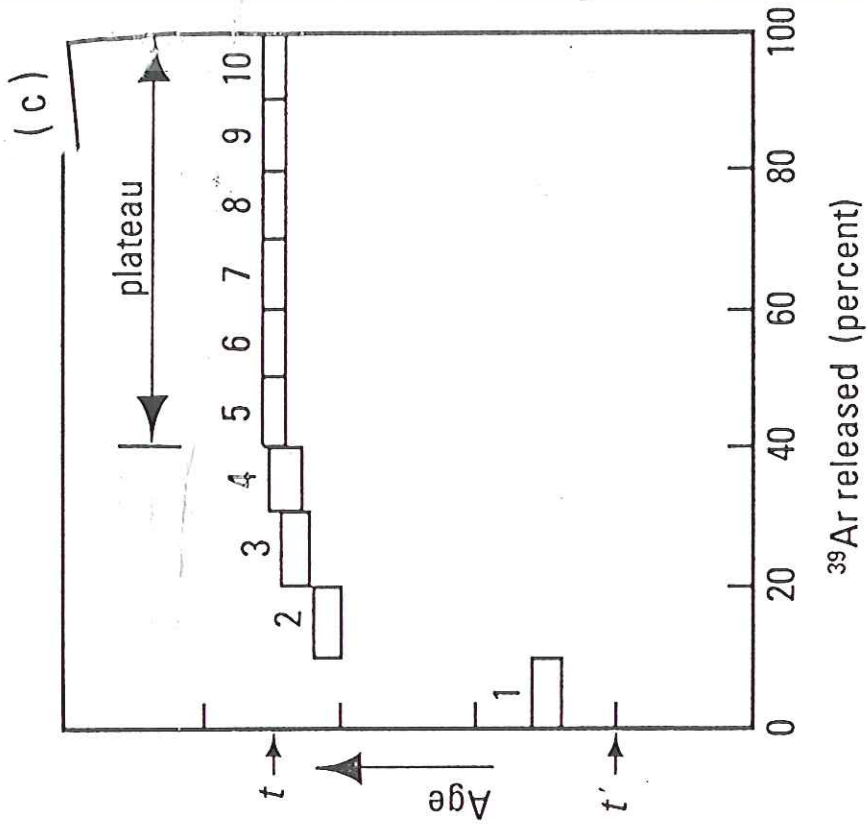
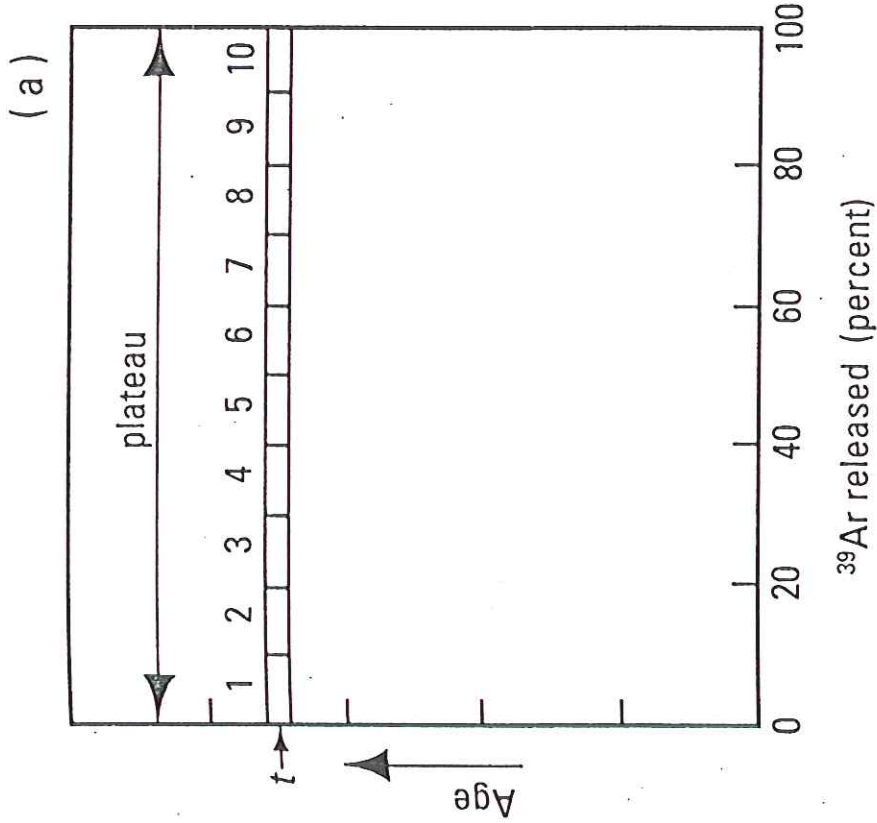


Fig. 3.11. (a) Hypothetical $^{40}\text{Ar}/^{39}\text{Ar}$ Ar age spectrum for an undisturbed sample. Because the sample has been a closed system since its formation, each of the gas increments (1 through 10) gives the same age, t . Increment 1 is released at the lowest temperature and 10 at the highest.

(c) Age spectrum for a hypothetical sample of age t that lost 15% of its Ar when it was heated at time t' . The increments (5 through 10) released at high temperatures still give the formation age and form a "plateau."

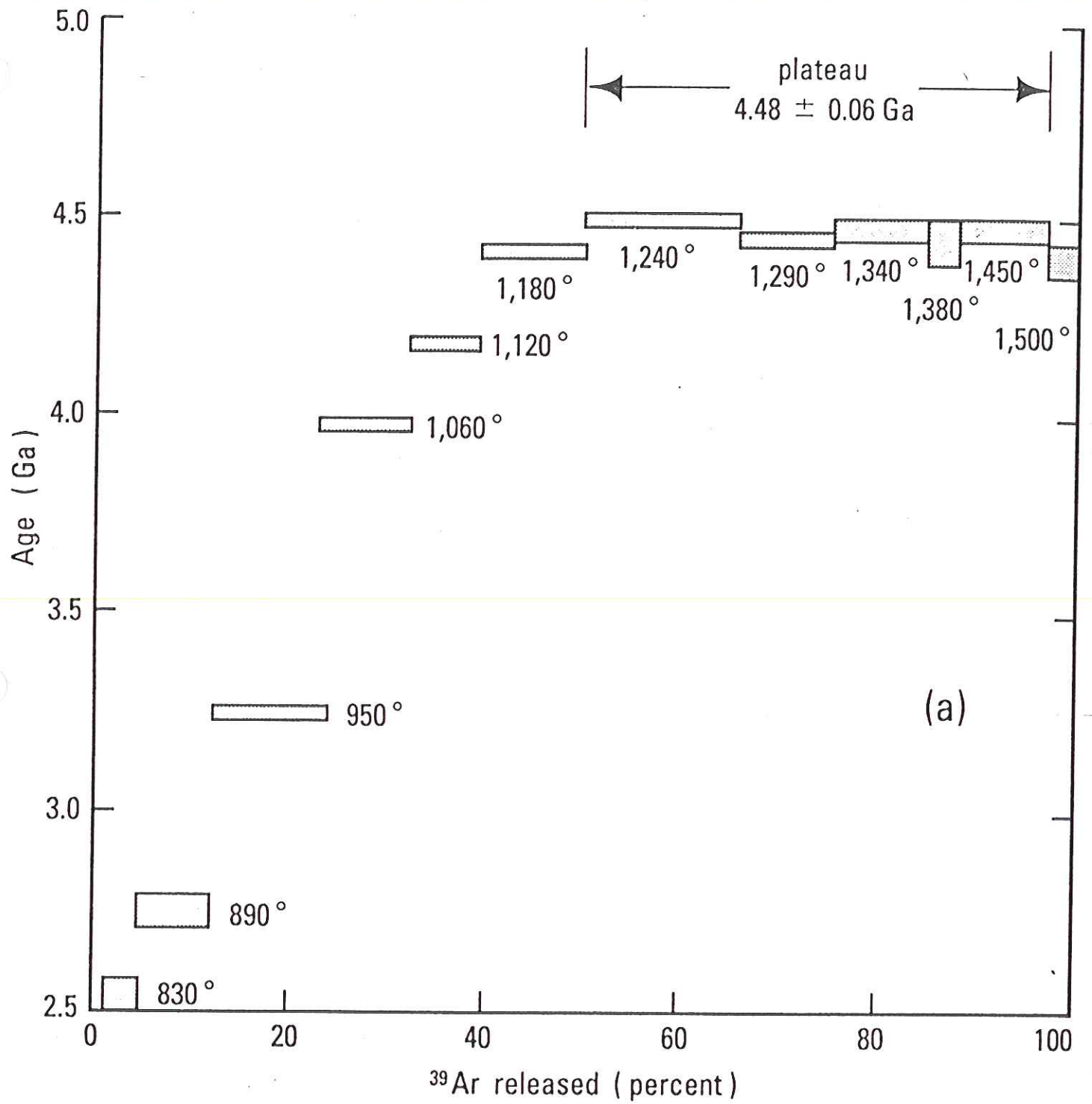


Fig. 3.12. $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum for the meteorite Menow, which lost 25% of its Ar at about 2.5 Ga. The temperature at which each gas increment was released is shown in degrees Celsius. (After Turner, Enright, and Cadogan, 1978.)

SINGLE SAMPLE POTASSIUM-ARGON AGES USING THE OMEGATRON

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*Department of Geodesy and Geophysics,
 University of Cambridge, England*

Received 13 April 1966

The principal experimental limitations of the potassium-argon method of age determination by conventional mass spectrometry are: (i) weight of sample required, (ii) mass discrimination effects, (iii) inaccuracies of potassium determination by flame photometry at low concentrations, (iv) inaccuracies arising from the potassium and argon measurements being made upon different portions of the bulk sample. Some of the advantages of the Omegatron method have already been reported [1]. Experiments showed that under normal operating conditions the instrument was free from mass discrimination effects and that the problem of sample weight used in argon determinations could be resolved, but as long as the potassium content is measured by flame photometry, the difficulty of separating a large sample still exists. A solution to this latter problem and to that of overcoming inaccuracies introduced by inhomogeneity within the bulk sample has been suggested by Merrigue [2].

³⁹K is produced when a sample containing ³⁹K is irradiated by fast neutrons. ⁴⁰Ar is present in the sample due to the natural decay of the isotope ⁴⁰K.

For a sample of age t :

$$^{40}\text{Ar} = ^{40}\text{K}(e^{t/\tau} - 1), \quad (1)$$

where ⁴⁰K, ⁴⁰Ar denote the amounts of the isotopes at present and $\tau = 1/\text{total decay constant for } ^{40}\text{K}$. The amount of ³⁹Ar produced when this sample is irradiated is given by:

$$^{39}\text{Ar} = ^{39}\text{K} \int \varphi(E) \sigma(E) dE, \quad (2)$$

where $\varphi(E)$ denotes the neutron flux at energy E and $\sigma(E)$ the cross-section of ³⁹K for such neutrons. The integration is performed over all incident neutron energies. It follows that:

$$^{40}\text{Ar}/^{39}\text{Ar} = ^{40}\text{K}/^{39}\text{K} \frac{(e^{t/\tau} - 1)}{\int \varphi(E) \sigma(E) dE}. \quad (3)$$

Let

$$J = I \int \varphi(E) \sigma(E) dE,$$

where I is the relative abundance of the potassium isotopes ³⁹K/⁴⁰K. J is then a measure of the neutron absorption; it is a constant at a given point in the can in which the sample is irradiated. Since from eqs. (3) and (4):

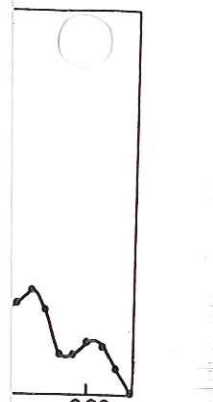
$$J = \frac{e^{t/\tau} - 1}{^{40}\text{Ar}/^{39}\text{Ar}},$$

it can be found from measurements on samples of known age. This argument assumes that no argon isotope is destroyed or produced in appreciable quantity by any other process during the irradiation. A study of the extensive literature suggests that this condition is satisfied.

Samples of known potassium-argon ages, enclosed within quartz phials, were arranged with a cadmium-shielded reactor can and irradiated with a total integrated flux of about 10^{19} neutrons per cm². The conditions were such that there was a neutron flux gradient over the length of the can. The relative abundances of the isotopes ⁴⁰Ar, ³⁹Ar and ³⁶Ar were afterwards determined using a standard argon extraction line in conjunction with the Omegatron. The quantity τ was calculated for each sample (correction having been made for the atmospheric contamination using the ³⁶Ar abundance). The results are shown in table 1.

In fig. 1 the values of J are shown plotted against sample position in the can. As would be expected from the known reactor characteristics, a continuous variation in J is obtained. The value of J at any point in the can may be interpolated from this curve. Thus, the inclusion of samples of unknown age into such a calibrated can and the subsequent measurement of their ⁴⁰Ar/³⁹Ar ratios enables their ages to be determined.

This method of potassium-argon age determi-



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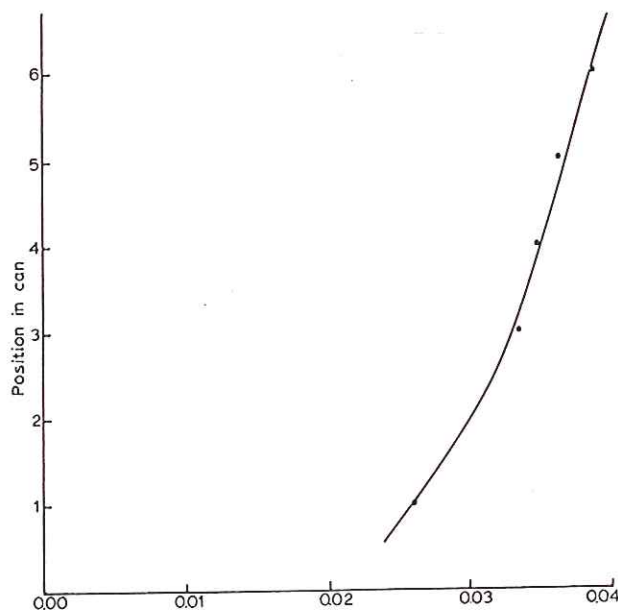
age
 V. E. Scott,
 scription and analysis,
 1947) 93.

Table 1

V/M = volume of radiogenic ^{40}Ar per g (mm^3 at N.T.P.). Age = potassium-argon age in million years as determined by conventional methods in this laboratory. Decay constants used: $\lambda_e = 0.584 \times 10^{-10} \text{ yr}^{-1}$, $\lambda_\beta = 4.72 \times 10^{-10} \text{ yr}^{-1}$.

Sample	Position in can	K ₂ O (%)	V/M	Age	$^{40}\text{Ar}/^{39}\text{Ar}$	J
Muscovite	1	10.19	0.0284	81.3	1.65	0.0261
Blank Phial	2	-	-	-	nil	-
Muscovite	3	10.25	0.1906	492	8.86	0.0335
Biotite	4	8.65	0.1615	436	7.44	0.0348
Muscovite	5	10.25	0.1906	492	8.19	0.0363
Muscovite	6	10.19	0.0284	81.3	1.11	0.0388

Fig. 1. Variation of J with position in reactor can. The can was 23 cm long, the units of position in the figure are approximately 3 cm.



nation has the following advantages: (i) the sample weight is orders of magnitude smaller than that required for a complete determination by conventional methods; (ii) a knowledge of the precise weight of the sample is not required and (iii) inhomogeneities in the distribution of potassium within the sample are of no consequence.

The authors would like to acknowledge the arrangement of irradiation facilities by Mr. D. H. Brown of A. W. R. E., Aldermaston.

References

- [1] R. L. Crasty and J. A. Miller, The Omegatron, a useful tool for argon isotope studies, *Nature* 207 (1965) 1146.
- [2] C. M. Merrihue, Trace element determinations and potassium-argon dating by mass spectrometry of neutron irradiated samples, *Trans. Am. Geophys. Union* 46 (1965) 125.

K-Ar good for volcanic
 rocks — data sediments
 by bracketing —
 example for
 J&R

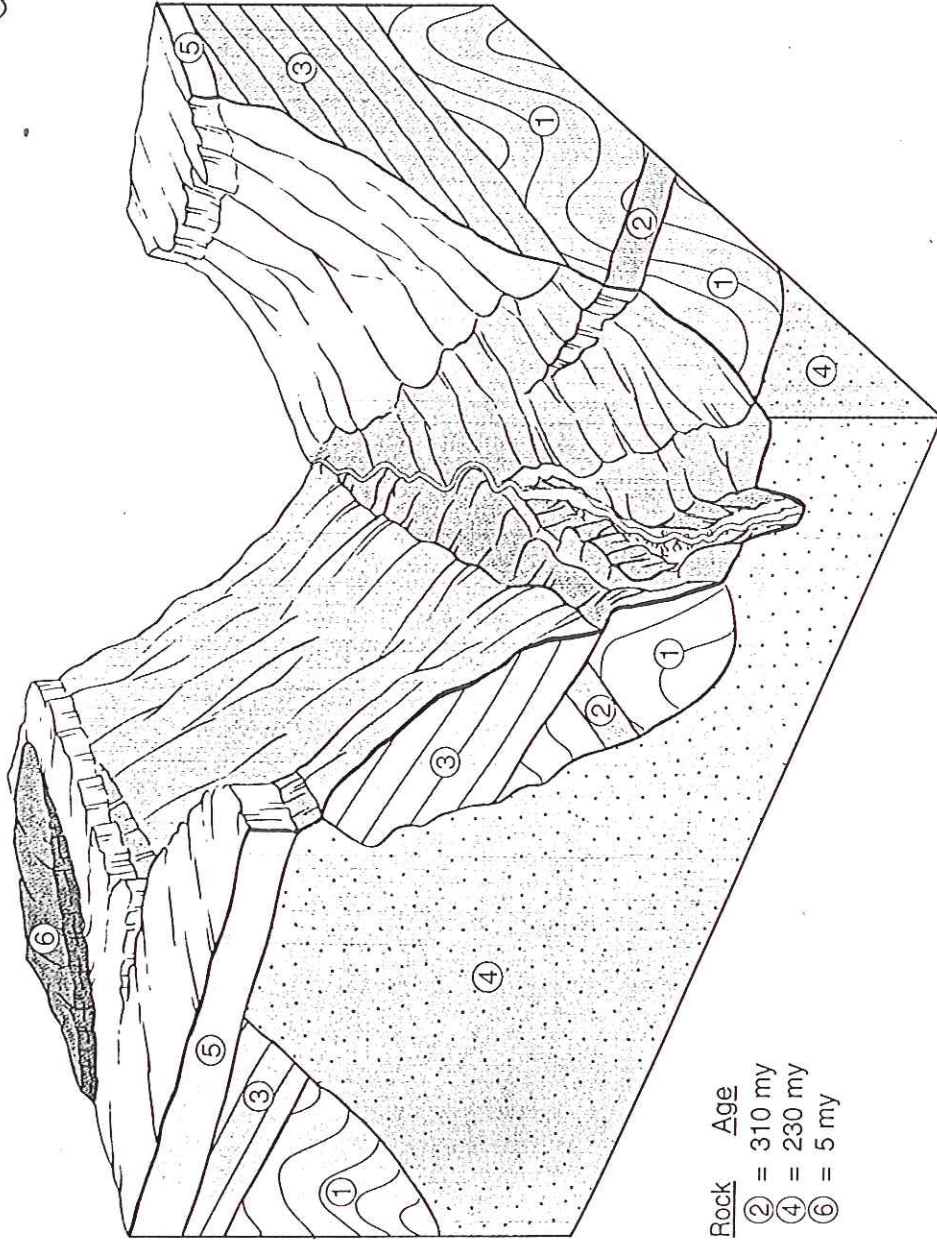
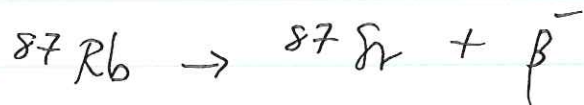


FIGURE 7.19

Radiometric dates obtained on igneous rocks can be fitted into the geologic time scale based on sedimentary rocks. This is done by bracketing the sedimentary rocks with the igneous rocks. In some instances radiometric dates can be assigned directly to the sedimentary rocks. See text for discussion.

Rb - Sr method :



see Fig. 3.4

$$t_{1/2} = 49 \text{ b.y.}$$

Useful for very old rocks that may have had complex histories

Rb is $Z=37$ just below K in periodic table

		^{84}Sr	0.5%
		^{86}Sr	10%
^{85}Rb	72%	^{87}Sr	7% 7% + daughters
^{87}Rb	28%	^{88}Sr	82.5%

~~ionic radii~~ Crystal abundance Fig I.9

$K = 10^7$, $Rb = 10^{3.5}$ by mass
— measured in parts per 10^9

$$\frac{K}{Rb} \text{ in atoms} = \frac{\frac{1}{2} \cdot 10^7}{10^{3.5}} = 0.5 \cdot 10^{3.5}$$

$$= 1600$$

Ionic radii are comparable — Fig. 49

$$K - 1.33 \text{ \AA}$$

$$Rb - 1.5 \text{ \AA}$$

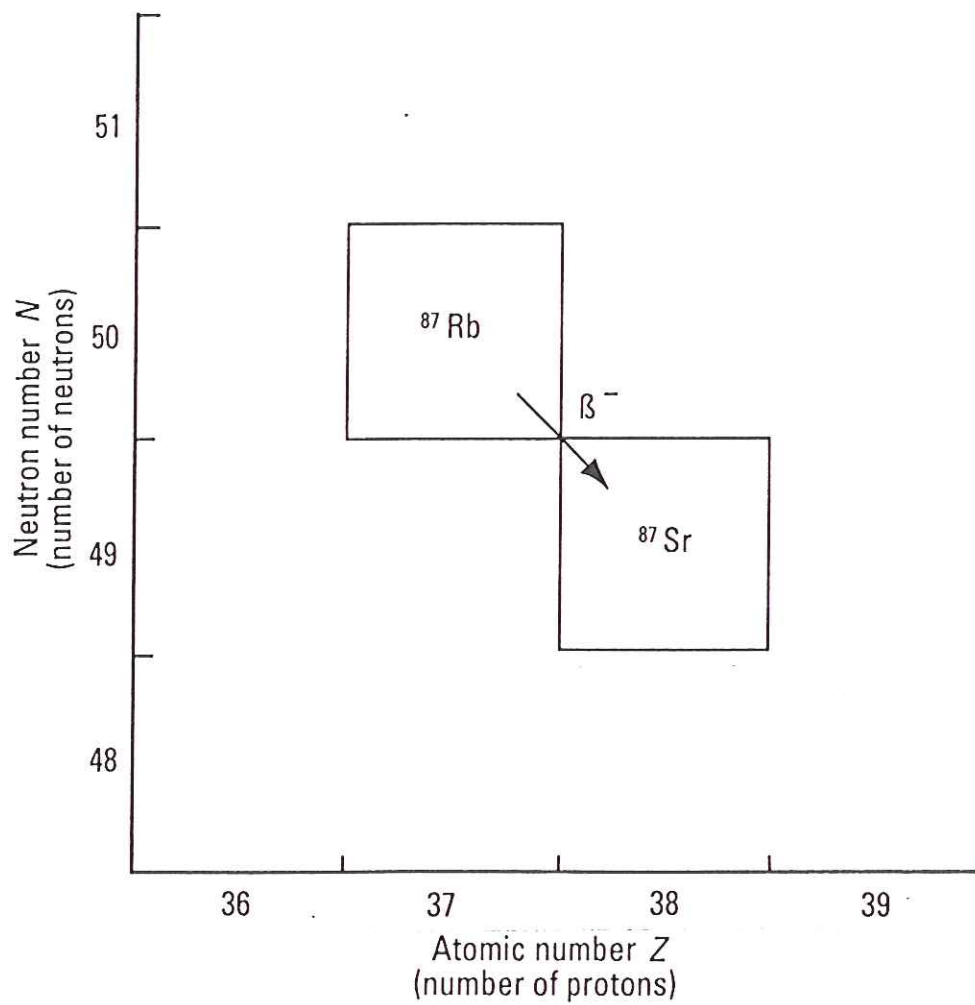


Fig. 3.4. ^{87}Rb decays to ^{87}Sr by β^- emission.

PERIODIC TABLE OF THE ELEMENTS

Strong tendency for outermost electrons to be lost to make full outer shell

H	Li	Na	K	Rb	Cs	Fr
1 Hydrogen 1.01	3 Lithium 6.94	11 Sodium 22.99	19 Potassium 39.10	37 Rubidium 85.47	55 Cesium 132.91	87 Francium (223)

Be	Mg	Ca	Sr	Ba	Ra
4 Beryllium 9.01	12 Magnesium 24.31	20 Calcium 40.08	38 Strontium 87.62	56 Barium 137.33	88 Radium 226.03

B	Al	Ga	In	Tl	Pb	Bi	Po	At	Rn
5 Boron 10.81	13 Aluminum 26.98	31 Gallium 69.72	49 Indium 114.82	81 Thallium 204.37	82 Lead 207.2	83 Bismuth 208.98	84 Polonium (209)	85 Astatine (210)	86 Radon (222)

C	Si	Ge	Sn	Pb
6 Carbon 12.01	14 Silicon 28.09	32 Germanium 72.59	50 Tin 118.69	82 Lead 207.2

N	P	As	Sb	Bi
7 Nitrogen 14.01	15 Phosphorus 30.97	33 Arsenic 74.92	51 Antimony 121.75	83 Bismuth 208.98

O	S	Se	Te	Po
8 Oxygen 16.00	16 Sulfur 32.06	34 Selenium 78.96	52 Tellurium 127.60	84 Polonium (209)

F	Cl	Br	I	Xe
9 Fluorine 19.00	17 Chlorine 35.45	35 Bromine 79.90	53 Iodine 126.90	54 Xenon 131.30

Ne	Ar	Kr	Xe	Rn
10 Neon 20.18	18 Argon 39.95	36 Krypton 83.80	54 Xenon 131.30	86 Radon (222)

Noble gases: outer shells filled; no tendency to gain or lose electrons

He	Ne	Ar	Kr	Xe	Rn
2 Helium 4.00	10 Neon 20.18	18 Argon 39.95	36 Krypton 83.80	54 Xenon 131.30	86 Radon (222)

Strong tendency to gain electrons to make full outer shell

O	S	Se	Te	Po
8 Oxygen 16.00	16 Sulfur 32.06	34 Selenium 78.96	52 Tellurium 127.60	84 Polonium (209)

N	P	As	Sb	Bi
7 Nitrogen 14.01	15 Phosphorus 30.97	33 Arsenic 74.92	51 Antimony 121.75	83 Bismuth 208.98

C	Si	Ge	Sn	Pb
6 Carbon 12.01	14 Silicon 28.09	32 Germanium 72.59	50 Tin 118.69	82 Lead 207.2

B	Al	Ga	In	Tl
5 Boron 10.81	13 Aluminum 26.98	31 Gallium 69.72	49 Indium 114.82	81 Thallium 204.37

Li	Na	K	Rb	Cs	Fr
3 Lithium 6.94	11 Sodium 22.99	19 Potassium 39.10	37 Rubidium 85.47	55 Cesium 132.91	87 Francium (223)

Be	Mg	Ca	Sr	Ba	Ra
4 Beryllium 9.01	12 Magnesium 24.31	20 Calcium 40.08	38 Strontium 87.62	56 Barium 137.33	88 Radium 226.03

Transition elements: valence electrons not in outer shell

Zn	Cu	Ni	Co	Fe	Mn	Cr	V	Ti	Sc
30 Zinc 65.38	29 Copper 63.55	28 Nickel 58.70	27 Cobalt 58.93	26 Iron 55.85	25 Manganese 54.94	24 Chromium 52.00	23 Vanadium 50.94	22 Titanium 47.90	21 Scandium 44.96

Cd	Ag	Pd	Rh	Ru	Tc	Mo	Nb	Zr	Y
48 Cadmium 112.41	47 Silver 107.87	46 Palladium 106.4	45 Rhodium 102.91	44 Ruthenium 101.07	43 Technetium (98)	42 Molybdenum 95.94	41 Niobium 92.91	40 Zirconium 91.22	39 Yttrium 88.91

Hg	Au	Pt	Ir	Os	Re	W	Ta	Hf	Rf
80 Mercury 200.59	79 Gold 196.97	78 Platinum 195.09	77 Iridium 192.22	76 Osmium 190.2	75 Rhenium 186.21	74 Tungsten 183.85	73 Tantalum 180.95	72 Hafnium 178.49	104 Rutherfordium 261.11

Mt	Hs	Bh	Sg	Db	Rf
109 Meitnerium (266)	108 Hassium (264)	107 Bohrium (262)	106 Seaborgium (263)	105 Dubnium 262.11	104 Rutherfordium 261.11

IA IIA IIIB IVB VB VIB VIIIB VIIIIB IB IIB IIIA IVA VA VIA VIIA VIIIA

Lanthanide (Rare Earth) Elements

La	Ce	Pr	Nd	Pm	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu
*57 Lanthanum 138.91	58 Cerium 140.12	59 Praseodymium 140.91	60 Neodymium 144.24	61 Promethium (145)	62 Samarium 150.4	63 Europium 151.96	64 Gadolinium 157.25	65 Terbium 158.93	66 Dysprosium 162.50	67 Holmium 164.93	68 Erbium 167.26	69 Thulium 168.93	70 Ytterbium 173.04	71 Lutetium 174.97

Actinide Elements

Ac	Th	Pa	U	Np	Pu	Am	Cm	Bk	Cf	Es	Fm	Md	No	Lr
**89 Actinium 227.03	90 Thorium 232.04	91 Protactinium 231.04	92 Uranium 238.03	93 Neptunium 237.05	94 Plutonium (244)	95 Americium (243)	96 Curium (247)	97 Berkelium (247)	98 Californium (251)	99 Einsteinium (252)	100 Fermium (257)	101 Mendelevium (258)	102 Nobelium (259)	103 Lawrencium (260)

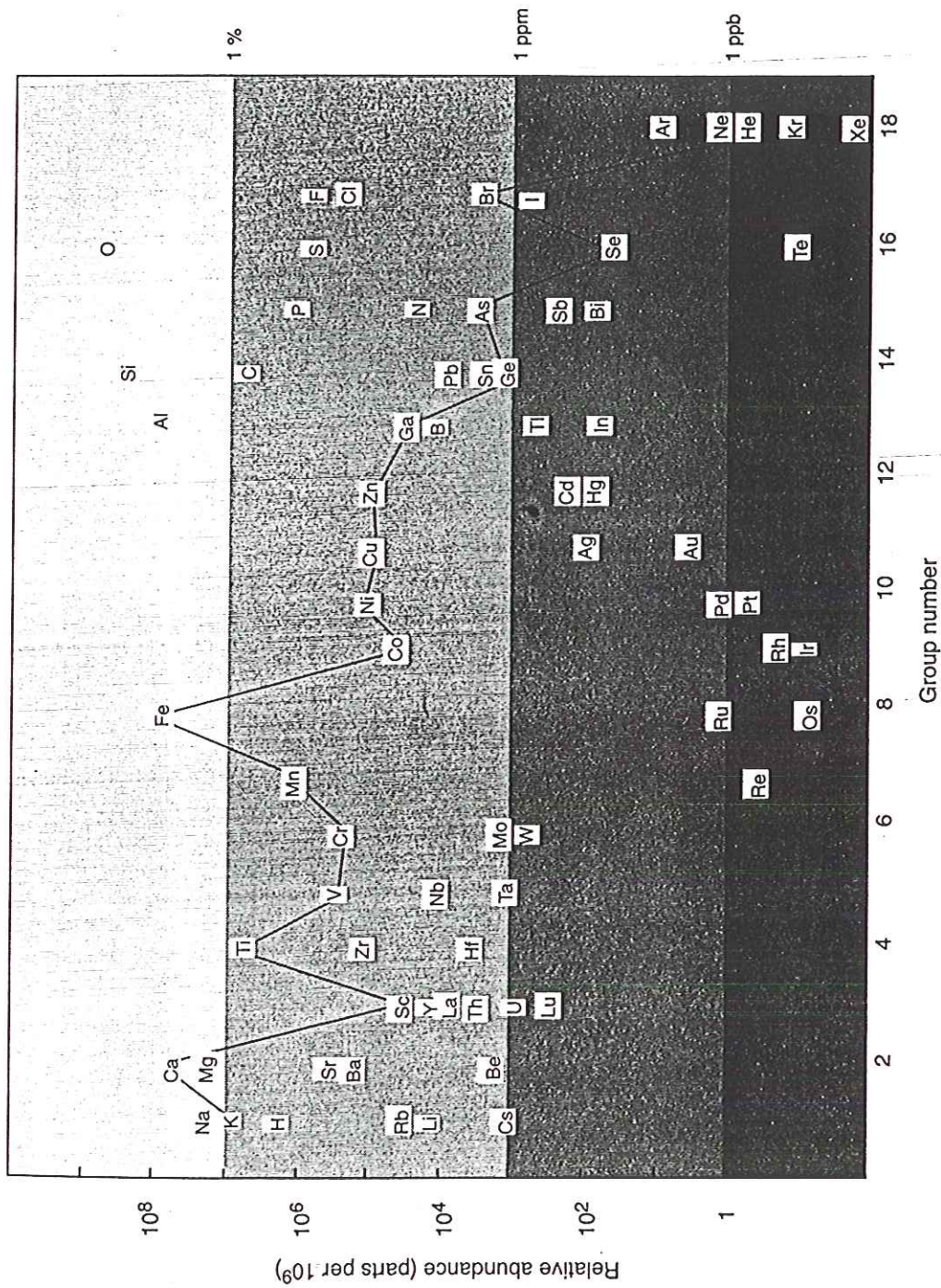


Fig. I.9 The abundance of elements in the Earth's crust, shown as the parts per billion (10^9) by mass on a logarithmic scale, with each division corresponding to a factor of 10 different from the neighbouring one. The horizontal scale is the group number in the periodic table. The line connects elements of the first long period (potassium to krypton) and is drawn as a guide to the eye. The different shading shows levels of abundance of one percent, one part per million (ppm) and one part per billion (ppb).

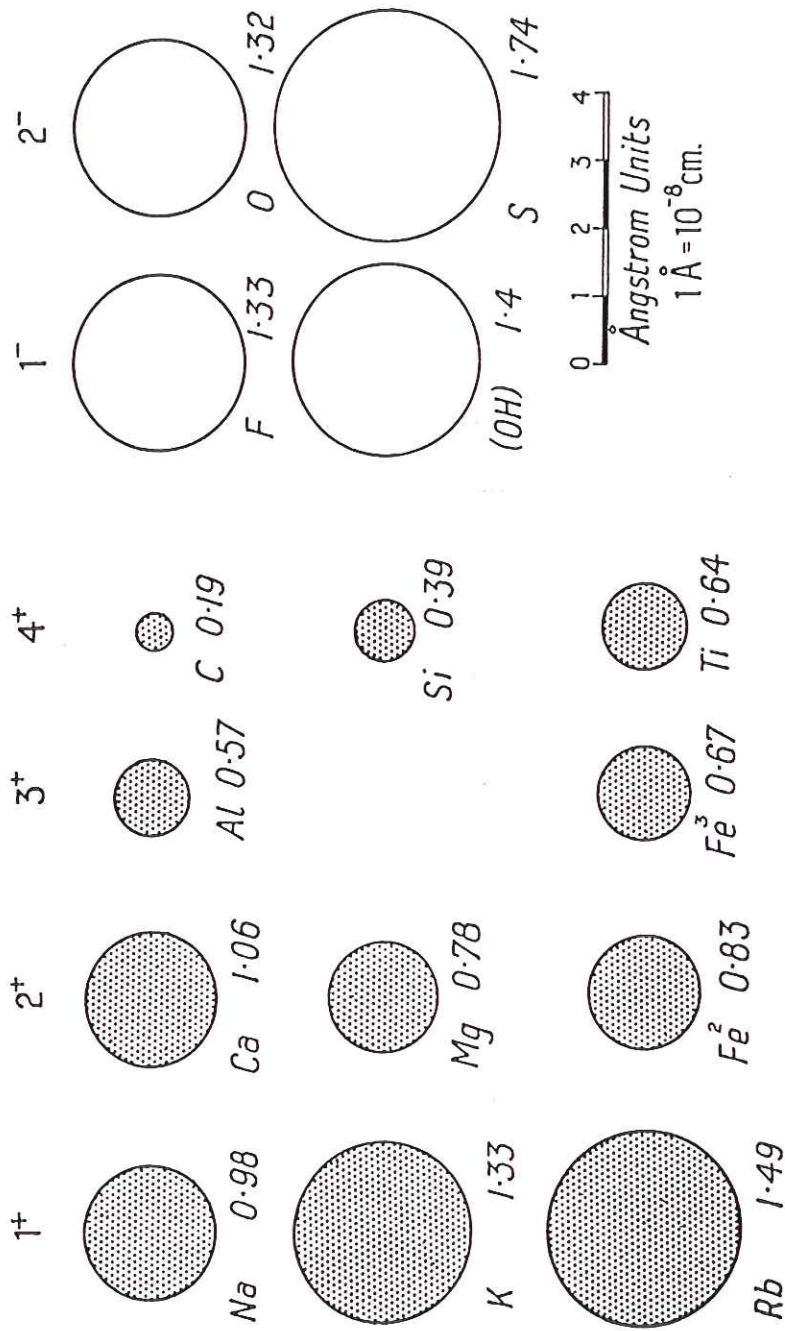


Fig. 49. Relative sizes of some positive ions (*cations*) and negative ions (*anions*) in crystals. The figure at the head of each column refers to the ionic charge or valency. The effective ionic radius of the field of influence of each ion within a crystal, conventionally regarded as spherical, is given in ångströms.

5

So in all potassium-bearing minerals, about ~~1000~~

$\frac{1}{\text{every } 1000}$ K atom is replaced by Rb

$\frac{1}{\text{every } 5000}$ replaced by ^{87}Rb

^{87}Sr in next column — also tightly bound in lattice — does not escape easily like Ar

Next explain concept of isochron diagram with aid of Fig 3.9

Separate 3 minerals, P, Q, R

Differing amounts of trace elements Rb, Sr

In (b) $t = t'$ ← length of arrows proportional to initial conc. of ^{87}Rb

Method utilizes fact that rocks composed of various minerals with varying amounts of Rb, Sr in each

Knowing only P', Q', R' tells us
no more than P, Q, R

But suppose we normalize by
dividing both axes by ^{86}Sr

~~Useful to use one near in mass~~

Useful to use one near in mass

Example — suppose $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right) = 2$

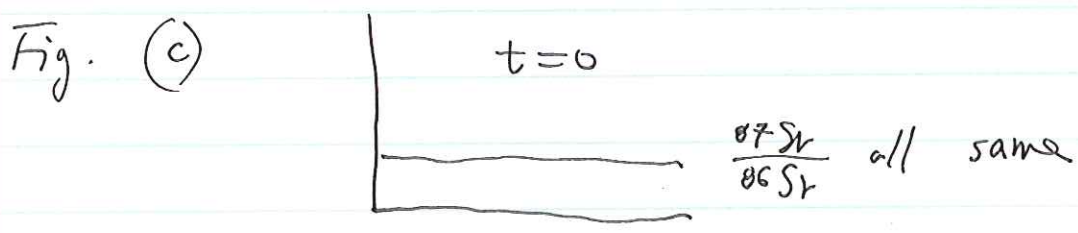
Why — because Sr incorporation into minerals is controlled by chemistry not mass

same for all samples

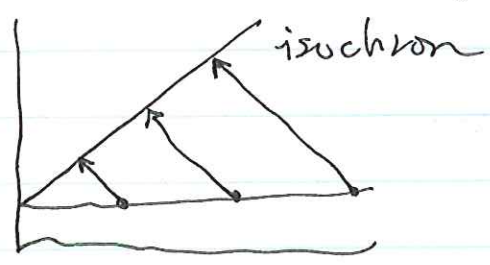
in fact it's more like $\frac{7}{10}$

but $\left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)$ different

NB — potentially confusing that numbers 1-14 are the same on Rb/Sr and ratio plots



Now, with time Fig. (d)



Line gets steeper with time

How do we determine the age?

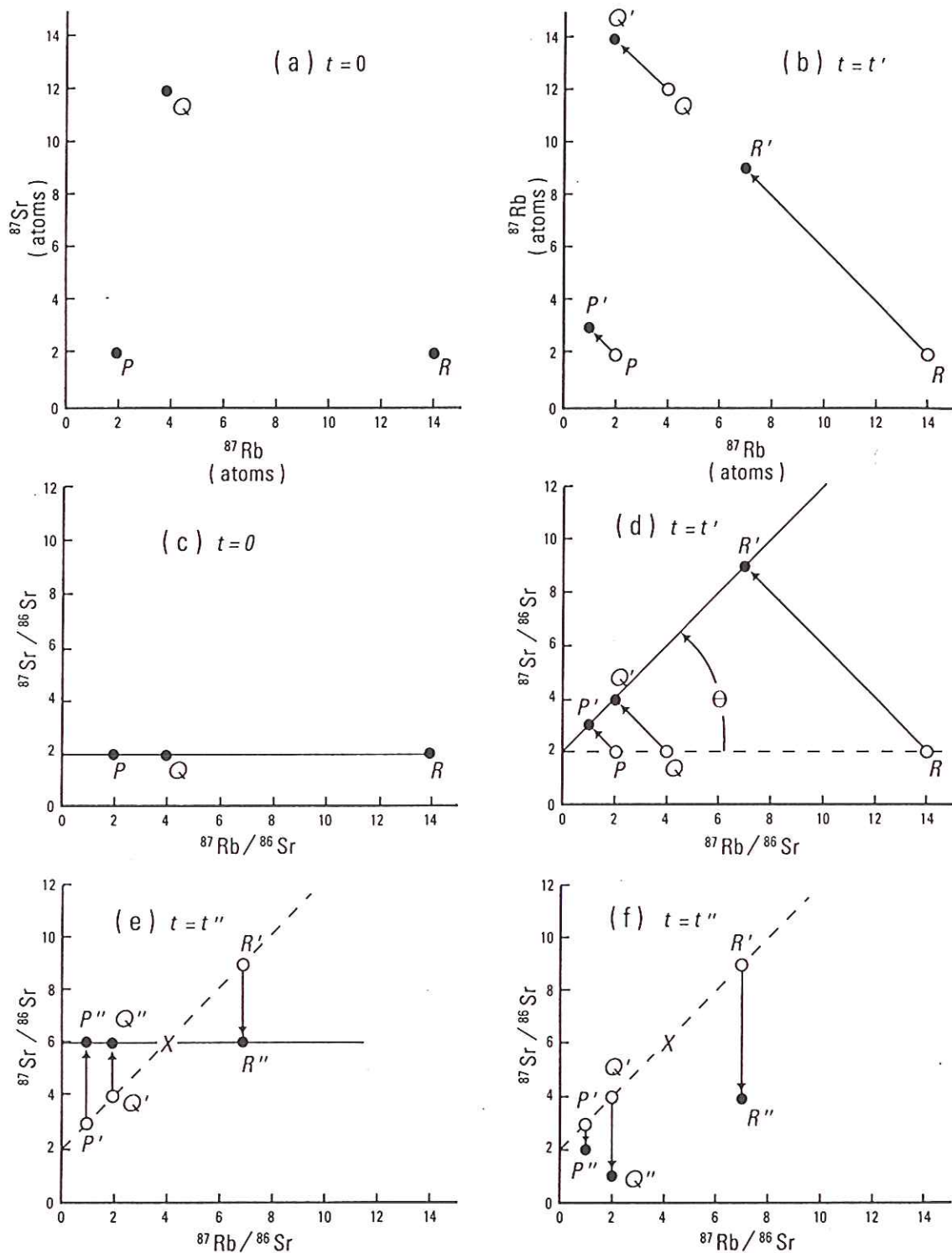
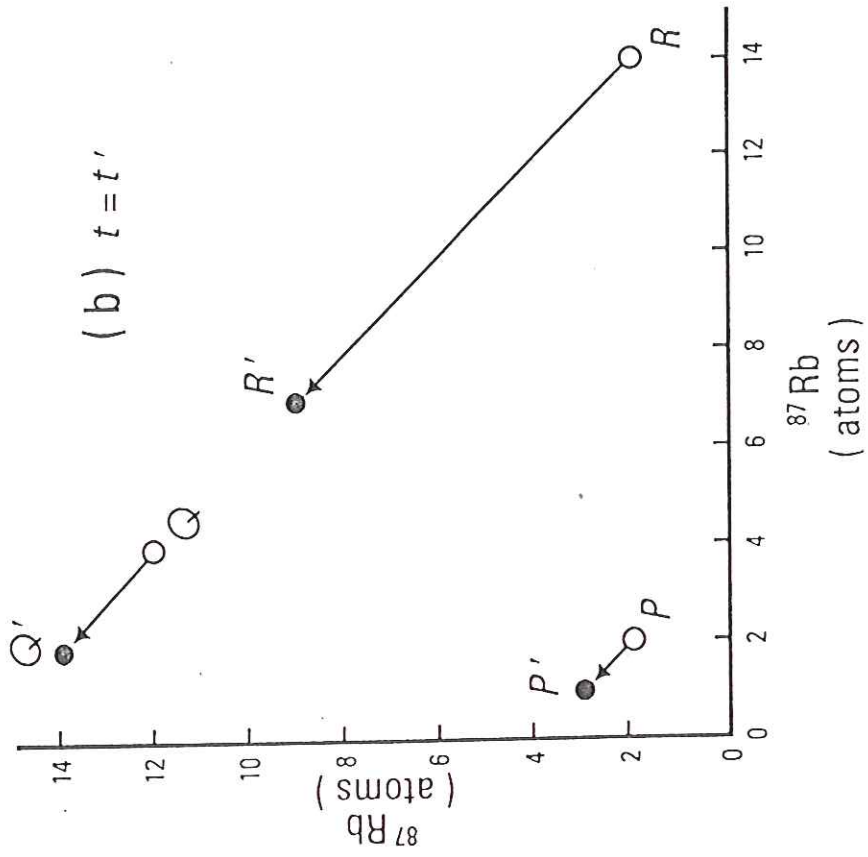
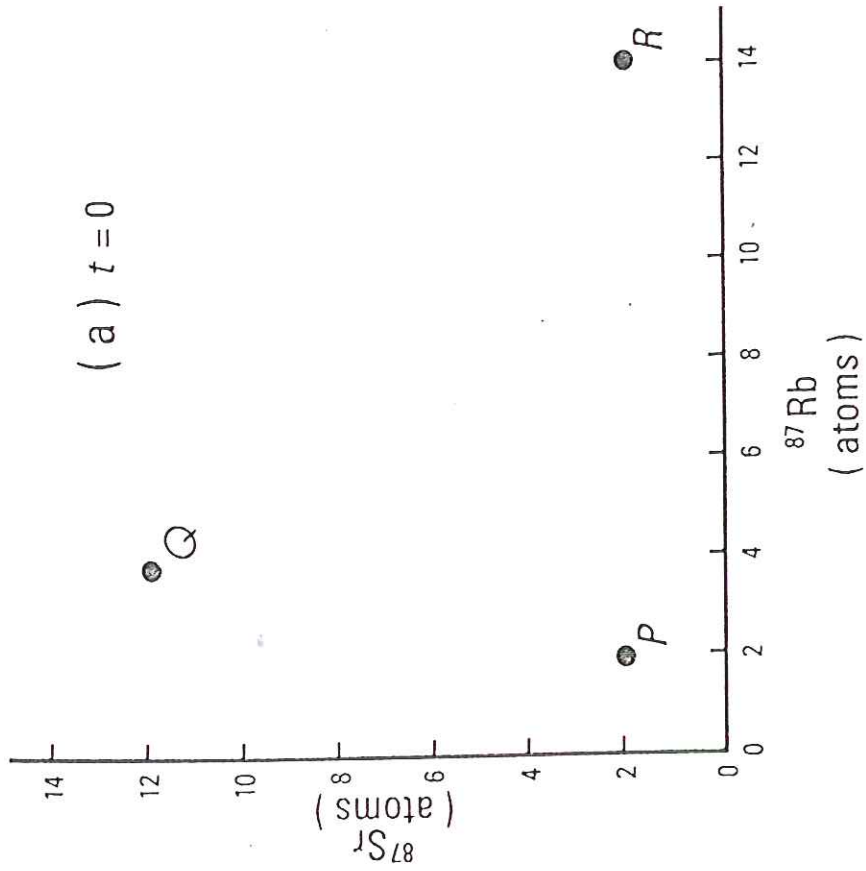


Fig. 3.9. (a) Plot of ^{87}Rb vs ^{87}Sr for three minerals, P, Q, and R, from a hypothetical rock of zero age. (b) Because of the decay of ^{87}Rb , points P, Q, and R move along trajectories of decreasing ^{87}Rb and increasing ^{87}Sr to P', Q', and R' after passage of time t' . The amount of movement is proportional to the ^{87}Rb content of the minerals, but this type of plot gives no information about the age of the rock. (c) The same data at $t = 0$ but normalized to ^{86}Sr . (d) After time t' has passed, the points still fall on a line, an isochron, whose slope is a function of age. (e) Complete resetting of the Rb-Sr clock at time t'' moves the points to a new "zero-age" isochron. The composition of the total rock is indicated by x. (f) Partial resetting, in this example due to loss of ^{87}Sr , results in the points scattering.

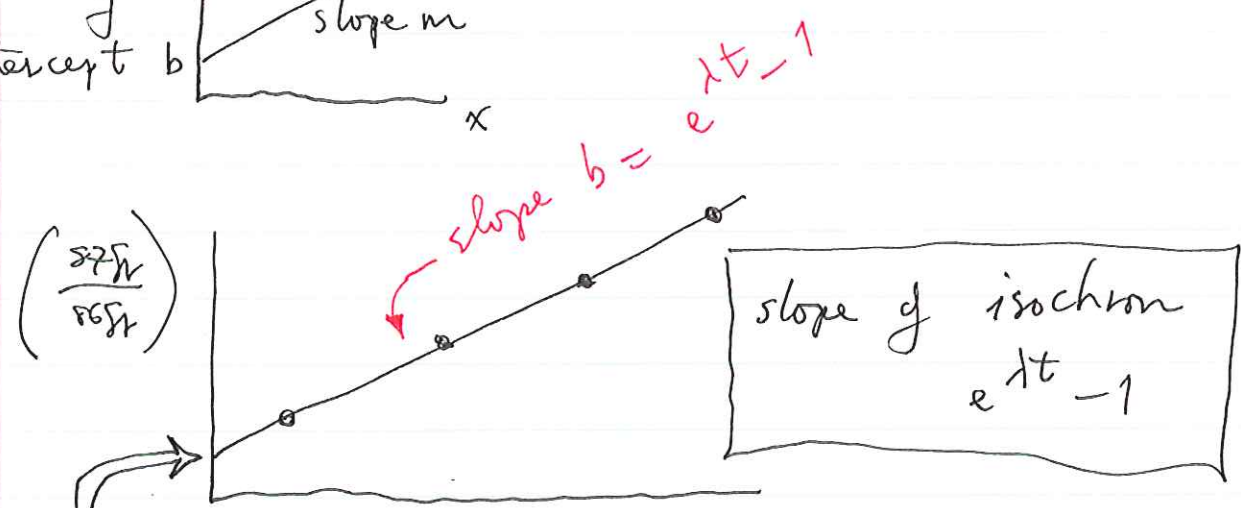
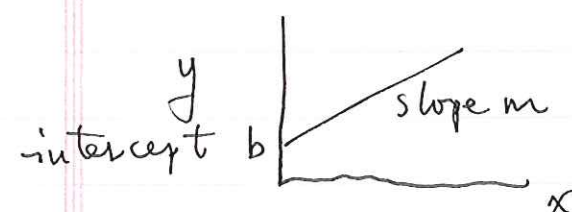


$$^{87}\text{Sr} = \left(^{87}\text{Sr} \right)_0 + \left(e^{\lambda t} - 1 \right) ^{87}\text{Rb}$$

↑
time zero

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_{\text{now}} = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_0 + \left(e^{\lambda t} - 1 \right) \left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}} \right)_{\text{now}}$$

Straight line $y = mx + b$

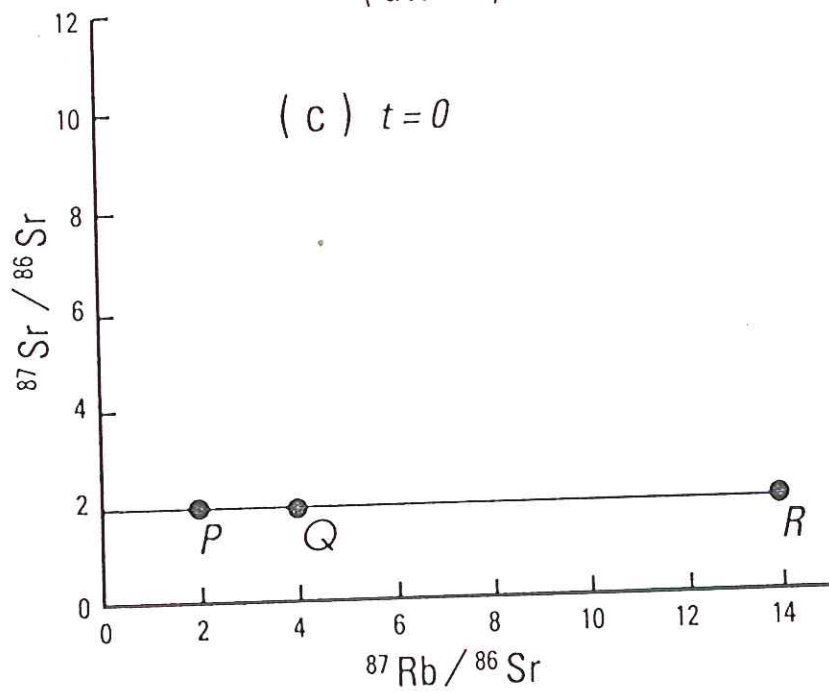
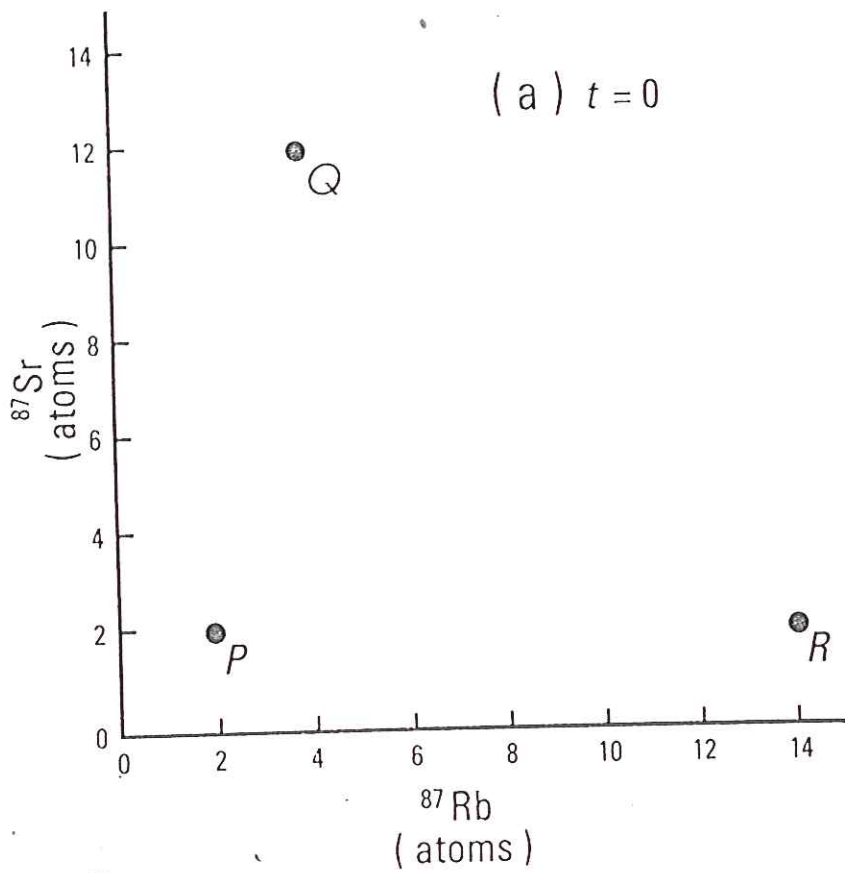


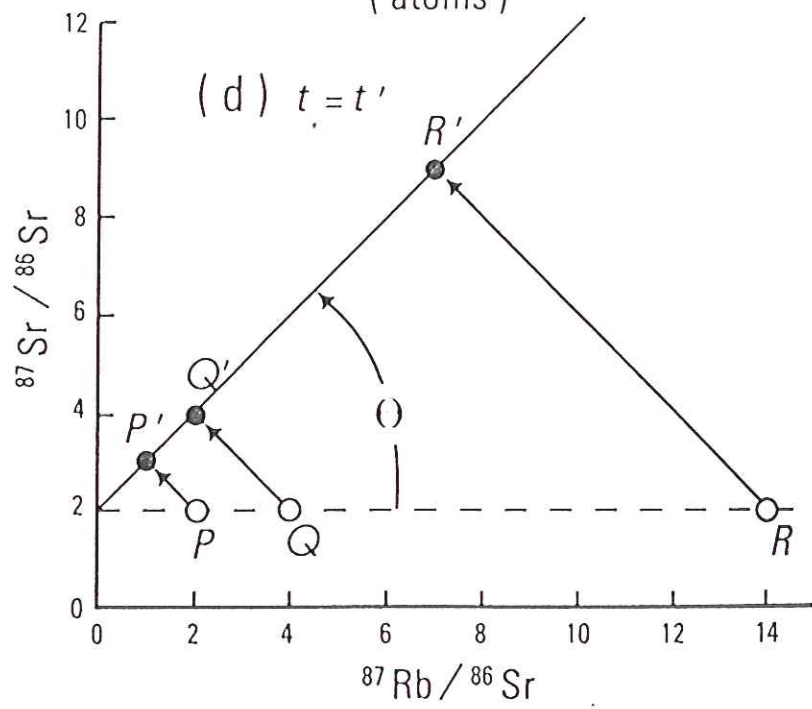
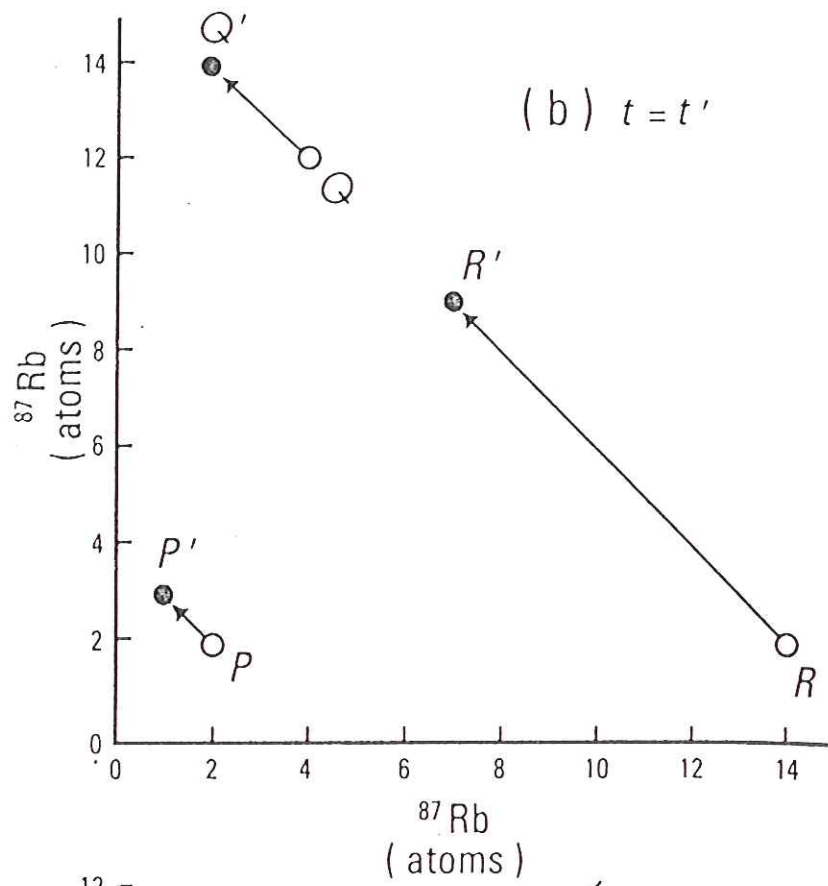
intercept
initial ~~ratio~~
strontium ratio
 $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_0 \approx 0.7$

i.e. this is Do - we determine it by using the isochron method!

Suppose now the rock is remelted & recrystallized at $t = t''$

The Sr isotopes will again homogenize in the melt and the ratio $^{87}\text{Sr}/^{86}\text{Sr}$ will again be ~~the same~~ the same in all x'tals





But the whole rock ratio will be higher because of the decay of $^{87}\text{Sr}/^{86}\text{Sr}$ during $0 \leq t \leq t''$

The process will start over — the slope will now give the new age of the rock — the previous age has been erased

Finally Fig. (f) shows partial resetting of the clock

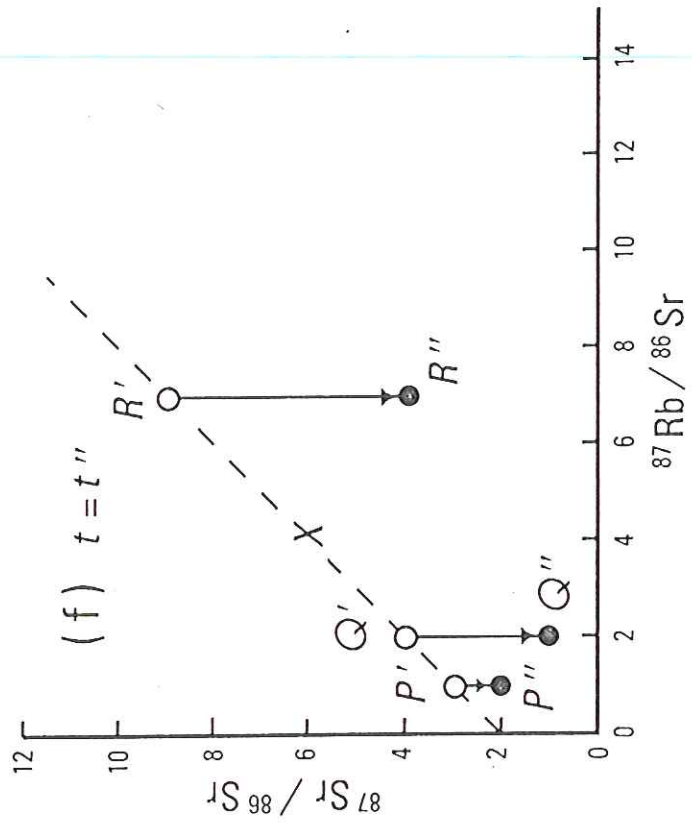
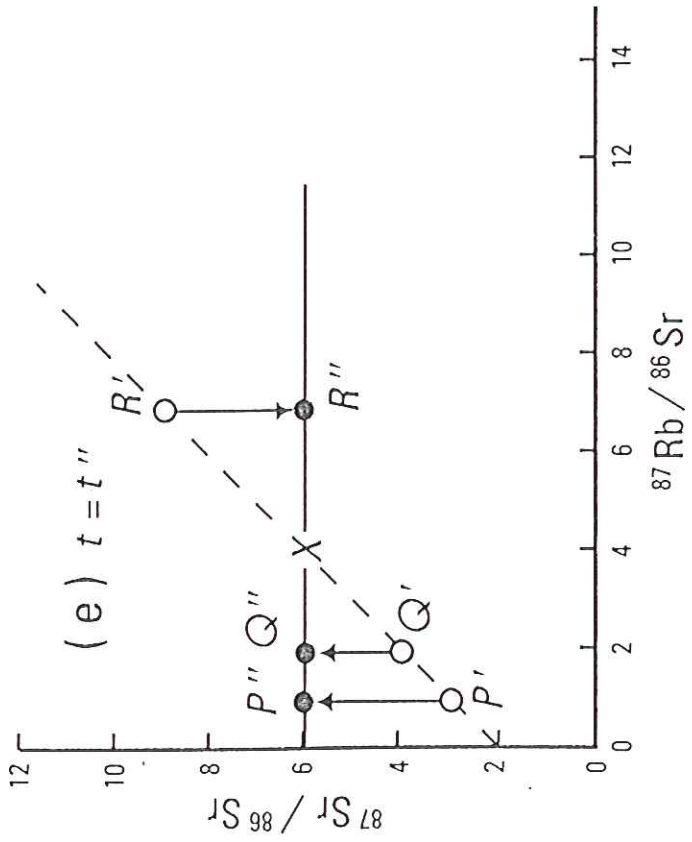
Mineral R has low Sr — not tightly bound — may lose more — mineral P may lose less — amount lost depends on how tightly bound in xtal

Such a disturbed rock — disturbed by a later heating event — can be recognized — will not form an isochron

Two advantages of method:

- solves initial daughter problem ($^{87}\text{Sr}/^{86}\text{Sr}$) is solved for as part of method
- tightness of isochron gives confidence in result — no partial resetting

Example
Fig 3.10



^{235}U in rocks of the Moon, and λ_2 the decay constants on ^{235}U can be solved algebraically by calculating a function of t until there is a good fit to the measured ratio. This method gives the most precise results. This will be discussed further in

and Pb loss the U-Th-Pb method with the aid of *isochron* or *concordia* diagrams to solve these problems and are the U-Th-Pb methods for which we discussed the isochron

grams

discussion, most of the simple methods have two requirements: (1) that the rock is closed, and (2) that the rock is homogeneous. But how can we find out if the system has been closed? These problems can be solved by using *isochron* diagrams. We will collectively call *isochron* diagrams their mathematical equivalent. They provide an exact measure of age for systems that have not been disturbed. They are especially useful for systems that are the basis of most of the work. They are important and they work.

isochron diagrams, each of which is the subject of this section, only determine the age of the

isochron, or *isotope evolution diagram*, for simplicity. The most comprehensive treatment was given in 1961 by L. O. Nicolaysen, Physical Research, Univer-

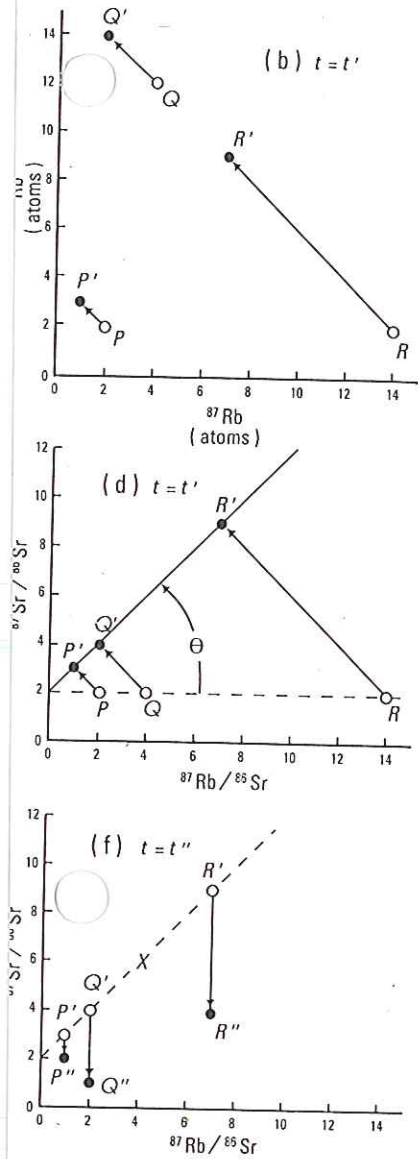
sity of Witwatersrand, South Africa, who applied it to Rb-Sr data and suggested that it could be used for U-Pb data as well (Nicolaysen, 1961). It is now a widely used geochronological tool, applicable in one form or another to all of the decay schemes used for radiometric dating.

The isochron method has two significant advantages over the simple accumulation clock. First, it circumvents the problem of the amount of the initial daughter. That information need not be known—it is one of the answers provided by the method. Second, the method is self-checking, providing the user with information about the degree to which the sample has behaved as a closed system.

As the name implies, an isochron is a line of equal time. It is obtained by analyzing several minerals from the same rock, or several rocks that formed from the same source at the same time but with differing amounts of the parent and daughter elements. On a simple graph, the amount of the parent isotope is plotted on the abscissa (x axis) and the amount of the daughter isotope is plotted on the ordinate (y axis), both values being normalized to (divided by) the amount of a nonradiogenic isotope of the same element as the daughter. If the samples have been closed systems since they formed, the points will fall on a line whose slope is a function of the age of the rock. The intercept of the line on the abscissa gives a measure of the initial daughter. At the moment, this description of the isochron may seem a bit cryptic, but the method is really quite simple, as the next few paragraphs will make clear.

The trick to the isochron diagram is the normalization of both parent and daughter isotopes to a third isotope of the daughter element. To see exactly what normalization does and how the isochron works, let us first consider what happens when the data are not normalized and consist solely of the amounts of the parent and daughter isotopes, using the Rb-Sr decay scheme to illustrate.

Suppose that we separate three minerals, P , Q , and R , from a newly formed rock ($t = 0$), determine their contents of ^{87}Rb and ^{87}Sr , and plot this information on a graph. For simplicity, imagine that the results can be expressed in small numbers of atoms. Mineral P is low in both Rb and Sr, mineral Q is low in Rb and high in Sr, and mineral R is high in Rb and low in Sr, so the graph looks like Figure 3.9a. At $t = 0$, the positions of the points on this graph depend only on the amounts of Rb and Sr in the minerals, i.e. they depend on the chemical composition of the minerals. If we reanalyze these minerals after letting them sit for some length of time, until their age is t' , the points on our graph will have moved because of the decay of ^{87}Rb to ^{87}Sr . In each sample, the decay of one atom of ^{87}Rb results in an increase of exactly one atom of ^{87}Sr , and so the points will have moved along tra-



erals, *P*, *Q*, and *R*, from a hypothetical
 b, points *P*, *Q*, and *R* move along trajec-
 to *P'*, *Q'*, and *R'* after passage of time t' .
 the ^{87}Rb content of the minerals, but this
 e of the rock. (c) The same data at $t = 0$
 passed, the points still fall on a line, an
 Complete resetting of the Rb-Sr clock at
 isochron. The composition of the total
 is example due to loss of ^{87}Sr , results in

jectories of 45° toward decreasing ^{87}Rb and increasing ^{87}Sr (Fig. 3.9b). Since the number of atoms of ^{87}Rb that decay in any given period of time is proportional to the number present, the distance the points move along the trajectories is a direct function of their ^{87}Rb content. Thus, if *P* decreases by one atom of ^{87}Rb , it will increase by one atom of ^{87}Sr and end up at *P'*. Point *Q* will move two units to *Q'* because it has twice the ^{87}Rb content of *P*. Point *R* has seven times as much ^{87}Rb as *P* so it will move seven times farther along its trajectory.

What information does a diagram like that in Figures 3.9a and 3.9b provide? Not much. If we had all six of the values shown in Figure 3.9b then we could calculate an age for the rock because we would know how far each point had moved along its trajectory, and that distance is a function of time. The fact is, however, that because we cannot determine *P*, *Q*, and *R*, only *P'*, *Q'*, and *R'*, we have no way of calculating the age, t' . In short, we would be stymied by the initial daughter problem.

Let us see what happens when these same hypothetical data are normalized. Normalization converts the isotopic data into ratios by measuring the amounts of the parent and daughter isotopes relative to that of another stable isotope of the daughter element, in this case Sr. Any of the available Sr isotopes could be used (Table 3.2), but there is some advantage in using an isotope as close as possible to the daughter in both mass and natural abundance, and by convention ^{86}Sr is universally used for normalization. For simplicity, let us assume that for every two atoms of ^{87}Sr in sample *P* there is one atom of ^{86}Sr . From Figure 3.9 we can see that there are two atoms of ^{87}Rb in sample *P*, so both ratios $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{87}\text{Rb}/^{86}\text{Sr}$ for sample *P* are equal to 2. If the ratio $^{87}\text{Sr}/^{86}\text{Sr}$ is 2 for sample *P*, then it must also be 2 for samples *Q* and *R*, so the data plot on a straight, horizontal line with $^{87}\text{Sr}/^{86}\text{Sr}$ equal to 2 for all values of $^{87}\text{Rb}/^{86}\text{Sr}$ (Fig. 3.9c). The value of $^{87}\text{Sr}/^{86}\text{Sr}$ when $^{87}\text{Rb}/^{86}\text{Sr}$ is zero, i.e. the intersection of the line on the ordinate, is the initial isotopic ratio of ^{87}Sr to ^{86}Sr in the rock—in this case 2.

Why are the Sr isotopic ratios the same even though the Rb and Sr contents of the minerals vary? The reason is that when a rock forms, all of the isotopes of any given element in the rock are homogenized. Consider, for example, the formation of a new rock by melting of an old rock. When the old rock is melted, all of the constituents, isotopes and elements, are thoroughly mixed by convection of the liquid and diffusion of the atoms throughout the melt. As the new rock begins to solidify, the minerals crystallize according to definite chemical rules that are governed by the composition, temperature, and pressure of the melt. Some minerals form early and extract from

the liquid those elements that constitute their crystals. This early crystallization changes the elemental composition of the remaining liquid, which leads to the formation of other minerals of different composition. Mineral species that form early in the process tend to incorporate more Sr than Rb, whereas those that form late tend to be high in Rb and low in Sr. Even though the relative amounts of Rb and Sr vary from mineral to mineral, however, the chemical processes of crystallization do not fractionate isotopes of the same element, so the isotopic composition of the Sr and Rb is the same in all of the minerals. This means that for a rock whose age is zero, the ratios of ^{87}Sr to ^{86}Sr in all of the minerals will be identical, whereas the ratios of ^{87}Rb to ^{86}Sr will vary from one mineral species to another.

As our hypothetical rock ages, the isotopic compositions of the samples P , Q , and R will move along their respective trajectories as before (Fig. 3.9d). Each decay of an atom of ^{87}Rb results in the addition of an atom of ^{87}Sr , so as the ratio $^{87}\text{Rb}/^{86}\text{Sr}$ decreases there is a corresponding increase in $^{87}\text{Sr}/^{86}\text{Sr}$, and the magnitude of change for any given period of time is a function of the Rb content, or in this case of the ratio $^{87}\text{Rb}/^{86}\text{Sr}$. With our normalized data, however, the points P' , Q' , and R' will always fall on a straight line whose slope is a direct function of the age of the rock, i.e. the older the rock, the steeper the slope of the line. Furthermore, this line will always intersect the ordinate at the value of the initial isotopic composition of Sr. This must be so because ^{87}Rb is zero at this intersection, so there can be no increase in ^{87}Sr over time. Thus, the isochron method gives both the age and the initial amount of daughter isotope in a rock solely from its current isotopic composition.

The equation for the isochron diagram is based on a simple modification of Equation 3.5. For the Rb–Sr system Equation 3.5 is

$$^{87}\text{Sr} = (^{87}\text{Sr})_0 + (e^{\lambda t} - 1)^{87}\text{Rb} \quad (3.14)$$

where ^{87}Sr and ^{87}Rb are the total amounts at time t and $(^{87}\text{Sr})_0$ is the amount of initial ^{87}Sr at $t = 0$. We can normalize the isotopic values in this equation by dividing all terms by the constant ^{86}Sr :

$$\frac{^{87}\text{Sr}}{^{86}\text{Sr}} = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_0 + (e^{\lambda t} - 1) \frac{^{87}\text{Rb}}{^{86}\text{Sr}} \quad (3.15)$$

This is the equation of a straight line of the form

$$y = b + mx \quad (3.16)$$

where the initial Sr ratio is the intercept, b , on the y axis at a value, x , of $^{87}\text{Rb}/^{86}\text{Sr} = 0$, and the term $(e^{\lambda t} - 1)$ is the slope, m .

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position of the remaining liquid,
r minerals of different composi-
in the process tend to incorpo-
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ion, so there can be no increase
method gives both the age and
in a rock solely from its current

diag is based on a simple
Rb-Sr system Equation 3.5 is

$$e^{\lambda t} - 1) \frac{{}^{87}\text{Rb}}{{}^{86}\text{Sr}} \quad (3.14)$$

nts at time t and $({}^{87}\text{Sr})_0$ is the
normalize the isotopic values in
the constant ${}^{86}\text{Sr}$:

$$e^{\lambda t} - 1) \frac{{}^{87}\text{Rb}}{{}^{86}\text{Sr}} \quad (3.15)$$

of the form

$$y = mx + b \quad (3.16)$$

at b , on the y axis at a value, x ,
is the slope, m .

When Equation 3.15 is solved for t and the appropriate value is used for the decay constant, we have

$$t = 7.042 \times 10^{10} \log_e \left[\frac{\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} - \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right)_0}{\frac{{}^{87}\text{Rb}}{{}^{86}\text{Sr}}} + 1 \right] \quad (3.17)$$

First, note the similarity of this equation to Equation 3.10. It is simply the Rb-Sr age equation with the isotope values expressed as ratios and a term for the initial ⁸⁷Sr added.

Second, note what happens if we rearrange Equation 3.15 slightly.

$$e^{\lambda t} - 1 = \left[\frac{\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} - \left(\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}}\right)_0}{\frac{{}^{87}\text{Rb}}{{}^{86}\text{Sr}}} \right] = \text{slope} \quad (3.18)$$

Now look again at equations 3.10 and 3.17 and Figure 3.9d. It should be obvious that the slope of the isochron is simply the net change in the ratio ⁸⁷Sr/⁸⁷Rb over time t . It may also be apparent that the tangent of the angle θ between the sloping isochron at $t = t'$ and the horizontal isochron at $t = 0$ is equal to ⁸⁷Sr/⁸⁷Rb, so that another way to express the relationship between age and slope is

$$t = 7.042 \times 10^{10} \log_e (\tan \theta + 1) \quad (3.19)$$

Next, let's see what happens to the isochron when the Rb-Sr clock is reset. Suppose that the rock is completely melted and allowed to recrystallize at some time t'' , just shortly after t' (Fig. 3.9e). Melting will rehomogenize the Sr isotopes, and the new minerals will once again share the same initial ratio of ⁸⁷Sr to ⁸⁶Sr, but that ratio will be higher than it was at $t = 0$ because of the decay of ⁸⁷Rb from time $t = 0$ to $t = t''$. Graphically, it is as if the isochron pivots about the isotopic composition of the total rock (point x in Figure 3.9e) and becomes horizontal again. The remelting and recrystallization have completely reset the Rb-Sr clock. A later age measurement will reflect the most recent time of recrystallization and "initial" Sr isotopic composition—the previous "age" and initial composition have been erased.

If the rock is not heated sufficiently to rehomogenize the Sr isotopes completely but enough so that Sr or Rb is allowed to move about, then the clock may not be completely reset. In such a *disturbed system*, there are a variety of things that can happen to the isotopic

composition of and within the rock. Either Sr or Rb isotopes, or both, may move in or out of the system, or the isotopes may simply be redistributed among the different minerals within the system. In such instances the results are unpredictable in detail, but the isotopic ratios for a system that has been disturbed almost invariably do not fall on any sort of isochron.

To see why, consider an oversimplified example that involves only the loss of ^{87}Sr from our hypothetical rock (Fig. 3.9f). Mineral *R* is high in Rb and low in Sr because a Rb atom is chemically and physically more compatible with the particular chemistry and crystal structure of that mineral than is a Sr atom. This means that a ^{87}Sr atom resulting from the decay of ^{87}Rb may find itself at a location within the crystal at which it is less firmly bound than either the original ^{87}Rb atom or a ^{86}Sr atom incorporated into the crystal when it formed. When the crystal is reheated, therefore, radiogenic ^{87}Sr atoms may be lost more easily than either ^{86}Sr or ^{87}Rb atoms. The same will not necessarily be true for mineral *P*, in which Sr is a more natural constituent. The ease with which radiogenic ^{87}Sr is lost from minerals *P*, *Q*, and *R* will be a function of many factors, such as chemical composition, crystal structure, and crystal defects, most of which are not directly related to the ratio $^{87}\text{Rb}/^{86}\text{Sr}$. Thus the movement of the compositions from *P'*, *Q'*, and *R'* to *P''*, *Q''*, and *R''* will not be an exact function of $^{87}\text{Rb}/^{86}\text{Sr}$ and so *P''*, *Q''*, and *R''* will not fall on a line except by a highly unlikely coincidence.

If we analyze the three minerals in our hypothetical rock and obtain data *P''*, *Q''*, and *R''*, we cannot determine exactly which isotopes have moved, where they have gone, or what amounts were involved. The data do, however, give us some very valuable information—they clearly reveal that the rock has been disturbed since it was formed, that the conditions of a closed system have been violated, that the Rb–Sr clock was partially reset at some unknown time, and that the age of the rock cannot be found from these particular data.

The effects of an incomplete resetting or disturbance are usually much more complicated than the simple example shown in Figure 3.9f. Isotopes may enter the system as well as leave, and there may be exchange of ^{87}Rb , ^{87}Sr , and ^{86}Sr between different minerals. These complications make it even less likely that the ratios will coincidentally fall on a straight line or isochron. For all practical purposes, the only way to move the isotopic compositions of samples from one isochron to another is by either radioactive decay through time or complete isotopic rehomogenization. Points that fall on an isochron, therefore, can confidently be interpreted as indicating the time of last isotopic homogenization, i.e. formation or reformation of the rock.

either ^{87}Rb or Rb isotopes, or both, the isotopes may simply be re-equilibrated within the system. In such cases, the isotopic ratios are not shown in detail, but the isotopic ratios almost invariably do not fall on a single line.

A simplified example that involves a hypothetical rock (Fig. 3.9f). Mineral Rb atoms are chemically and physically similar to Sr atoms. This means that a ^{87}Sr atom produced by the decay of ^{87}Rb at a location within the rock is indistinguishable from either the original ^{87}Rb or the original ^{87}Sr atoms. The same will not necessarily be true if ^{87}Sr is a more natural constituent of the rock. For example, if ^{87}Sr is lost from minerals P, Q, and R, such as chemical composition, then the movement of the common ^{87}Sr will not be an exact reflection of the ^{87}Rb decay. Thus, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio will not fall on a line except

if the system is closed. In our hypothetical rock and in natural rocks, we do not determine exactly which isotope was present at the time of formation. What amounts were in the system at the time of formation is some very valuable information. If the system has been disturbed since it was formed, the isotopic ratios have been violated. For example, if the system is reset at some unknown time, and the data are not consistent with the data obtained from these particular data. If the system is disturbed by melting or disturbance are usually not consistent with the data. A simple example shown in Figure 3.10a is well as leave, and there may be differences between different minerals. These differences may be such that the ratios will coincidentally fall on a single line. For all practical purposes, the isotopic ratios of samples from one isochron are indistinguishable. If the isotopic ratios of samples from one isochron decay through time or come from different locations that fall on an isochron, there is no way of knowing the time of last disturbance or reformation of the rock.

Thus, the isochron method is self-checking, providing not only the prospect of an age but also a statement on its validity.

An example of a valid Rb–Sr isochron is shown in Figure 3.10a in which data for the chondrite meteorite Tieschitz are plotted. Compare this isochron with the scattered data for the dikes and sills of the Precambrian Pahrump Group (Fig. 3.10b). For Tieschitz we can confidently conclude that its age is 4.52 Ga; for the data to fall on an isochron for any reason other than decay of ^{87}Rb within a closed system over time would be a highly improbable coincidence. For the Pahrump dikes and sills, however, we can only conclude that the samples are of different ages,¹² have not remained closed systems since their formation, or both. There is no way to determine the age of the rocks from these data, but neither do we face the prospect of calculating a Rb–Sr age that is incorrect and misleading. If we had made only a single analysis of any of the samples in Figure 3.10b our conclusions might have been quite different. It would have been necessary to make some estimate of the initial Sr composition and then calculate a simple accumulation age (sometimes called a *model age*) for the sample. The chances are great that the calculated age would have been incorrect, but, more seriously, we would have no way of knowing if it was. Thus, the isochron method, while more work because it requires multiple analyses, is worth the effort because it is self-checking.

The simple isochron can be used for virtually any of the decay schemes in Table 3.1. In practice, it is used extensively for the Sm–Nd, Lu–Hf, and Re–Os systems in exactly the way in which it is used for Rb–Sr dating, the only difference being the ratios expressed on the ordinate and abscissa (Table 3.3). The mathematics (Equations 3.14 through 3.19) are also identical, requiring only the substitutions of the proper isotope ratios (Table 3.2) and constants (Table 3.1). For the U–Pb and K–Ar systems, slightly different diagrams are used for reasons that we will examine.

The $^{40}\text{Ar}/^{39}\text{Ar}$ Age Spectrum and Isochron

The $^{40}\text{Ar}/^{39}\text{Ar}$ method is a form of K–Ar dating in which the sample is irradiated with *fast neutrons*¹³ in an atomic reactor to convert a fraction of the ^{39}K , which is the most common isotope of K (Table 3.2), to ^{39}Ar . The reaction of a fast neutron with a ^{39}K nucleus results in the addition of a neutron and the ejection of a proton, which

12. Whenever data are plotted on an isochron diagram, the initial assumption is that the samples (minerals or rocks) are of the same age and thus members of the same system. If this is not true, then the data will not fall on an isochron.

13. "Fast" neutrons are more energetic than "slow" neutrons. The terms are not precise ones, but energies above 0.02 MeV usually qualify neutrons as fast.

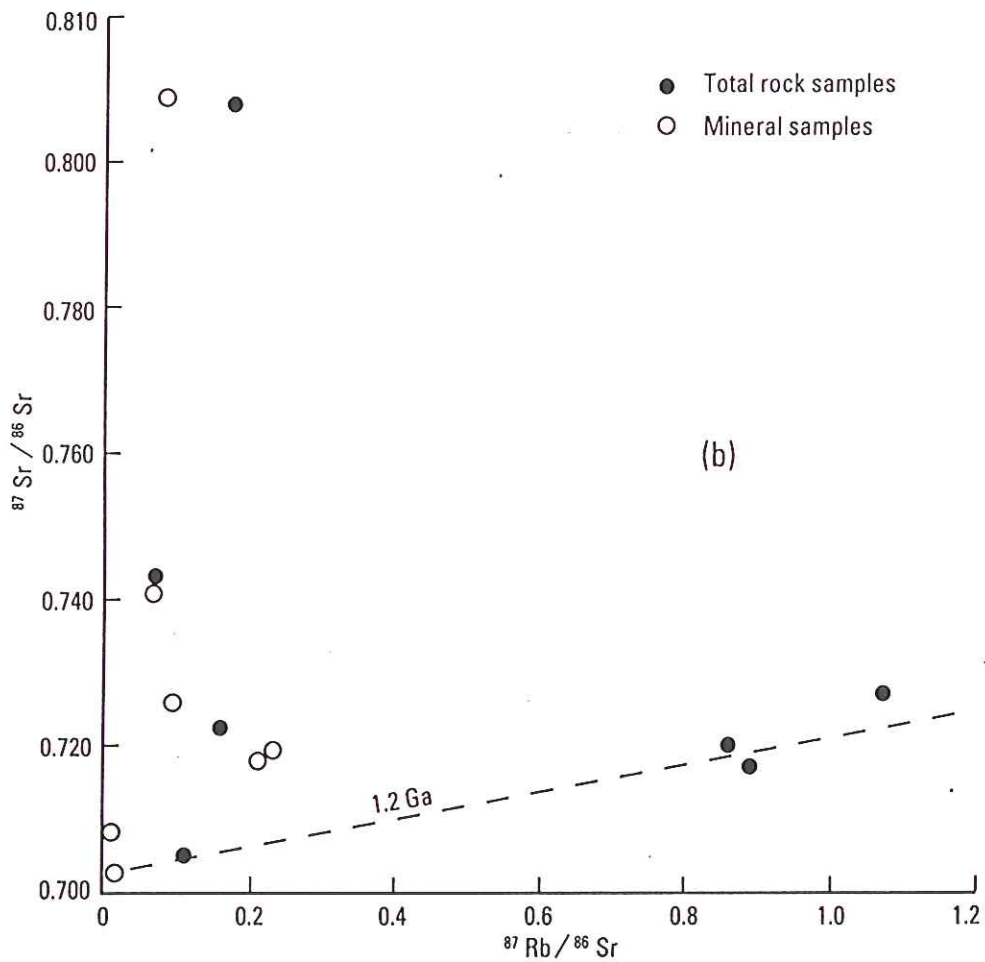
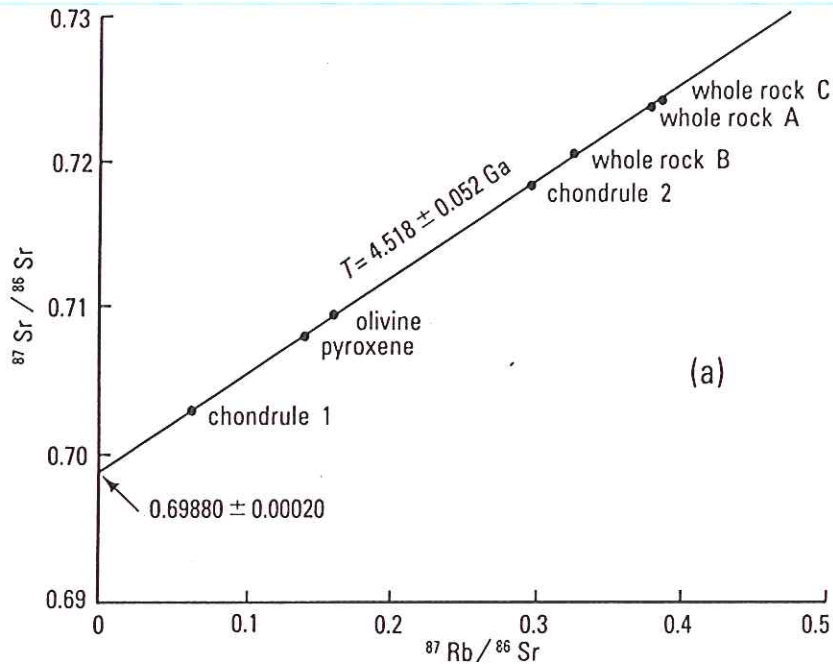
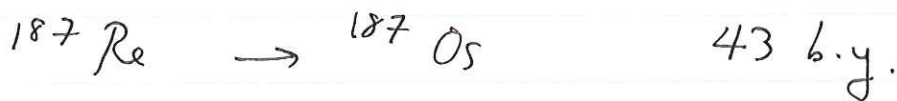
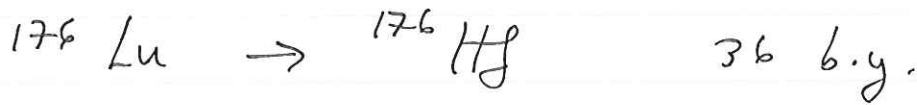
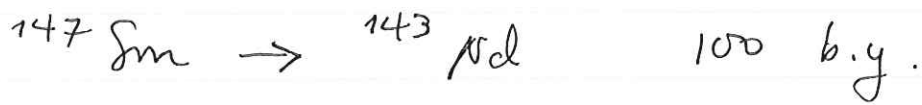


Fig. 3.10. (a) A mineral and whole-rock Rb-Sr isochron for the chondrite meteorite Tieschitz. (After Minster and Allègre, 1979a.) (b) Plot of Rb-Sr data for samples from igneous dikes and sills that intrude the Pahrump Group of the Panamint Mountains, California. The scatter of data shows clearly that these samples have been open systems, did not form at the same time, or both. Regardless, the ages of these rocks cannot be determined from these data. Other evidence indicates that these rocks are all about 1.2 Ga (dashed reference isochron). (After Wasserburg, Albee, and Lanphere, 1964.)

Isochron method can be applied to several other decay schemes



All are trace elements — but mass spectrometry is capable of measuring isotopic ratios using very small amounts

Sm & Lu — rare earths
 Re — a ~~heavy~~ heavy metal, as is Os

Show periodic table

Show mass ~~spec~~ spectrometer schematic — can be used to measure isotopic ratios with very high precision

Reason — both Faraday cups are measuring the same thing.

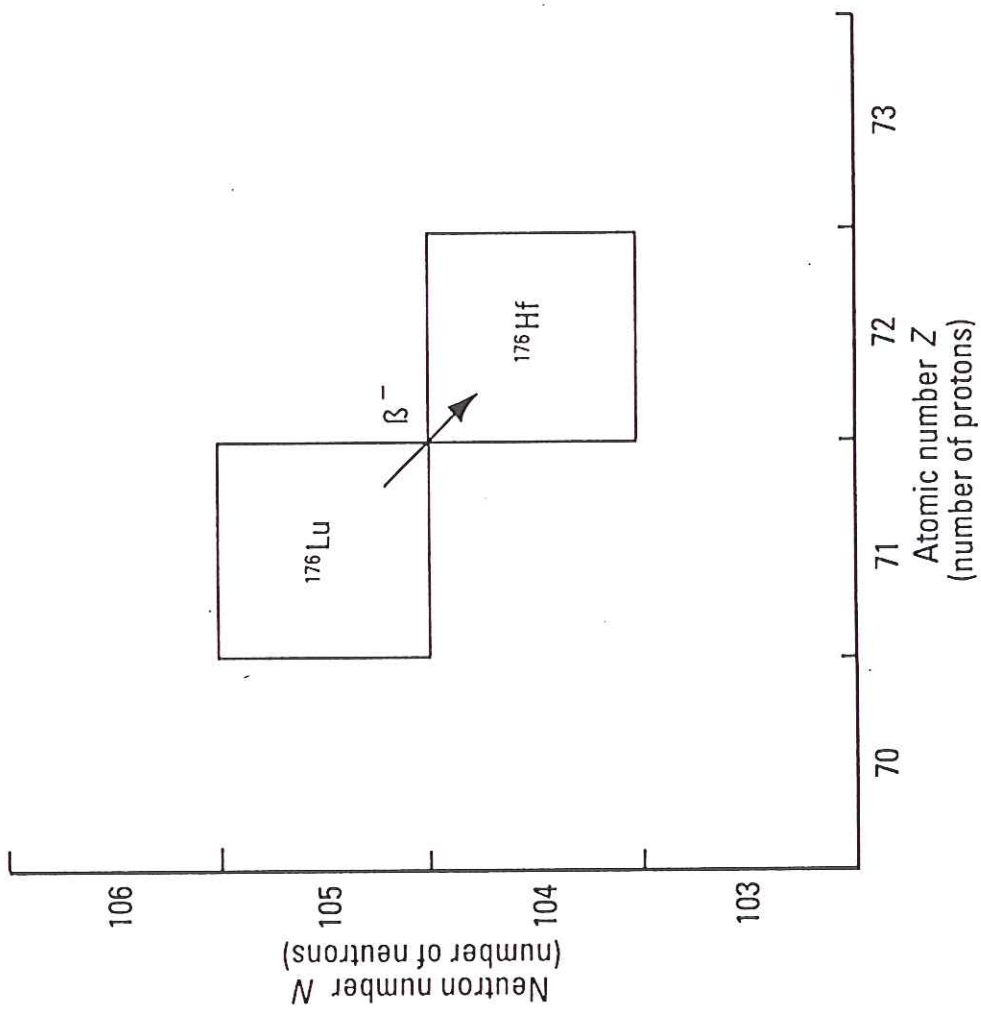


Fig. 3.6. ^{176}Lu decays to ^{176}Hf by β^- emission.

$$T_{1/2} = 39 \text{ b.y.}$$

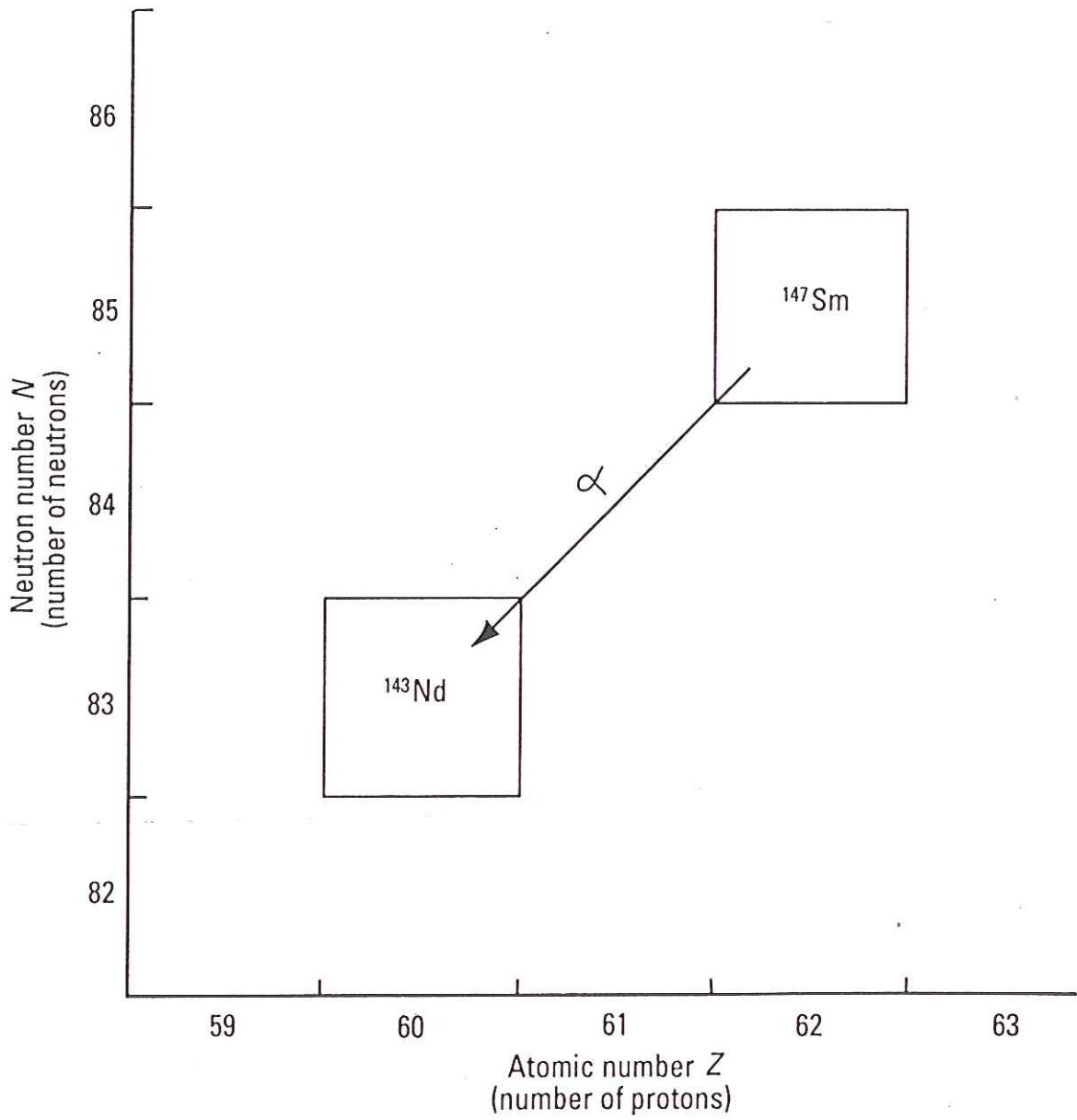


Fig. 3.5. ^{147}Sm decays to ^{143}Nd by α emission.

$$\tau_{1/2} = 100 \text{ b.y.}$$

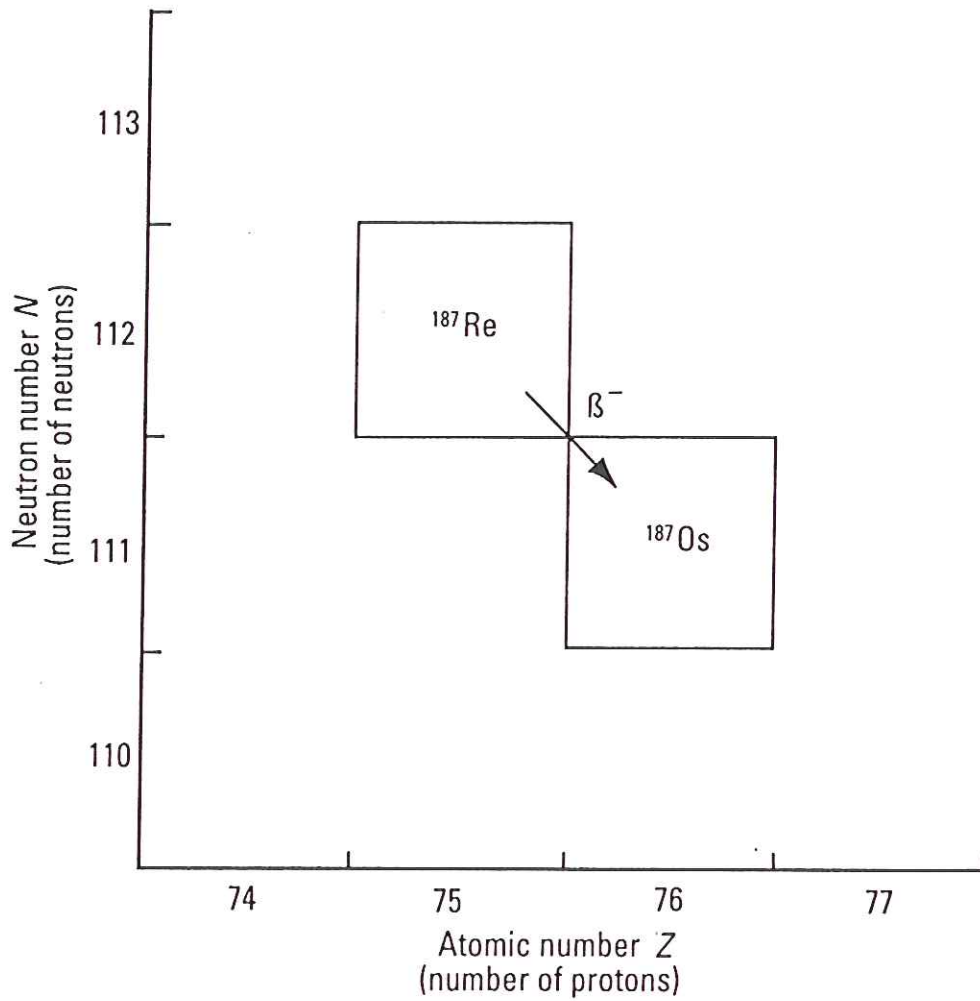


Fig. 3.7. ^{187}Re decays to ^{187}Os by β^- emission.

$$t_{1/2} = 50 \text{ b.y.}$$

TABLE 3.3
Parameters of the Common Radiometric Age-Diagnostic Diagrams

Diagram name	Ordinate (y axis)	Abscissa (x axis)
Rb-Sr isochron	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Rb}/^{86}\text{Sr}$
Sm-Nd isochron	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{147}\text{Sm}/^{144}\text{Nd}$
Lu-Hf isochron	$^{176}\text{Hf}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$
Re-Os isochron	$^{187}\text{Os}/^{186}\text{Os}$	$^{187}\text{Re}/^{186}\text{Os}$
$^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum	age or $^{40}\text{Ar}/^{39}\text{Ar}$	^{39}Ar released
$^{40}\text{Ar}/^{39}\text{Ar}$ isochron	$^{40}\text{Ar}/^{36}\text{Ar}$	$^{39}\text{Ar}/^{36}\text{Ar}$
Pb-Pb isochron	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{204}\text{Pb}$
U-Pb concordia	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$

PERIODIC TABLE OF THE ELEMENTS

Mn — Chemical Symbol
 25 — Atomic Number
 Manganese — Element Name
 54.94 — Atomic Weight

Strong tendency for outermost electrons to be lost to make full outer shell

H 1 Hydrogen 1.01	Li 3 Lithium 6.94	Be 4 Beryllium 9.01	Na 11 Sodium 22.99	Mg 12 Magnesium 24.31	K 19 Potassium 39.10	Ca 20 Calcium 40.08	Sc 21 Scandium 44.96	Y 39 Yttrium 88.91	Rb 37 Rubidium 85.47	Sr 38 Strontium 87.62	Cs 55 Cesium 132.91	Ba 56 Barium 137.33	Fr 87 Francium (223)	Ra 88 Radium 226.03
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B 5 Boron 10.81	Al 13 Aluminum 26.98	C 6 Carbon 12.01	Si 14 Silicon 28.09	N 7 Nitrogen 14.01	P 15 Phosphorus 30.97	O 8 Oxygen 16.00	S 16 Sulfur 32.06	F 9 Fluorine 19.00	He 2 Helium 4.00
B 5 Boron 10.81	Al 13 Aluminum 26.98	C 6 Carbon 12.01	Si 14 Silicon 28.09	N 7 Nitrogen 14.01	P 15 Phosphorus 30.97	O 8 Oxygen 16.00	S 16 Sulfur 32.06	F 9 Fluorine 19.00	He 2 Helium 4.00

Transition elements: valence electrons not in outer shell

Zn 30 Zinc 65.38	Cu 29 Copper 63.55	Ni 28 Nickel 58.70	Co 27 Cobalt 58.93	Fe 26 Iron 55.85	Mn 25 Manganese 54.94	Cr 24 Chromium 52.00	V 23 Vanadium 50.94	Ti 22 Titanium 47.90	Sc 21 Scandium 44.96
Zn 30 Zinc 65.38	Cu 29 Copper 63.55	Ni 28 Nickel 58.70	Co 27 Cobalt 58.93	Fe 26 Iron 55.85	Mn 25 Manganese 54.94	Cr 24 Chromium 52.00	V 23 Vanadium 50.94	Ti 22 Titanium 47.90	Sc 21 Scandium 44.96

La *57 Lanthanum 138.91	Ce 58 Cerium 140.12	Pr 59 Praseodymium 140.91	Nd 60 Neodymium 144.24	Pm 61 Promethium (145)	Sm 62 Samarium 150.4	Eu 63 Europium 151.96	Gd 64 Gadolinium 157.25	Tb 65 Terbium 158.93	Dy 66 Dysprosium 162.50	Ho 67 Holmium 164.93	Er 68 Erbium 167.26	Tm 69 Thulium 168.93	Yb 70 Ytterbium 173.04	Lu 71 Lutetium 174.97
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IA IIA IIIB IVB VIB VIIB VIIIB IIB IIIB IIIA IVA VA VIA VIIA VIIIA

Lanthanide (Rare Earth) Elements

La *57 Lanthanum 138.91	Ce 58 Cerium 140.12	Pr 59 Praseodymium 140.91	Nd 60 Neodymium 144.24	Pm 61 Promethium (145)	Sm 62 Samarium 150.4	Eu 63 Europium 151.96	Gd 64 Gadolinium 157.25	Tb 65 Terbium 158.93	Dy 66 Dysprosium 162.50	Ho 67 Holmium 164.93	Er 68 Erbium 167.26	Tm 69 Thulium 168.93	Yb 70 Ytterbium 173.04	Lu 71 Lutetium 174.97
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Actinide Elements

Ac **89 Actinium 227.03	Th 90 Thorium 232.04	Pa 91 Protactinium 231.04	U 92 Uranium 238.03	Np 93 Neptunium 237.05	Pu 94 Plutonium (244)	Am 95 Americium (243)	Cm 96 Curium (247)	Bk 97 Berkelium (247)	Cf 98 Californium (251)	Es 99 Einsteinium (252)	Fm 100 Fermium (257)	Md 101 Mendelevium (258)	No 102 Nobelium (259)	Lr 103 Lawrencium (260)
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Noble gases: outer shells filled; no tendency to gain or lose electrons

He 2 Helium 4.00	Ne 10 Neon 20.18	Ar 18 Argon 39.95	Kr 36 Krypton 83.80	Xe 54 Xenon 131.30	Rn 86 Radon (222)
----------------------------------	----------------------------------	-----------------------------------	-------------------------------------	------------------------------------	-----------------------------------

Li 3 Lithium 6.94	Be 4 Beryllium 9.01	B 5 Boron 10.81	C 6 Carbon 12.01	N 7 Nitrogen 14.01	O 8 Oxygen 16.00	F 9 Fluorine 19.00	Ne 10 Neon 20.18
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Li 3 Lithium 6.94	Be 4 Beryllium 9.01	B 5 Boron 10.81	C 6 Carbon 12.01	N 7 Nitrogen 14.01	O 8 Oxygen 16.00	F 9 Fluorine 19.00	Ne 10 Neon 20.18
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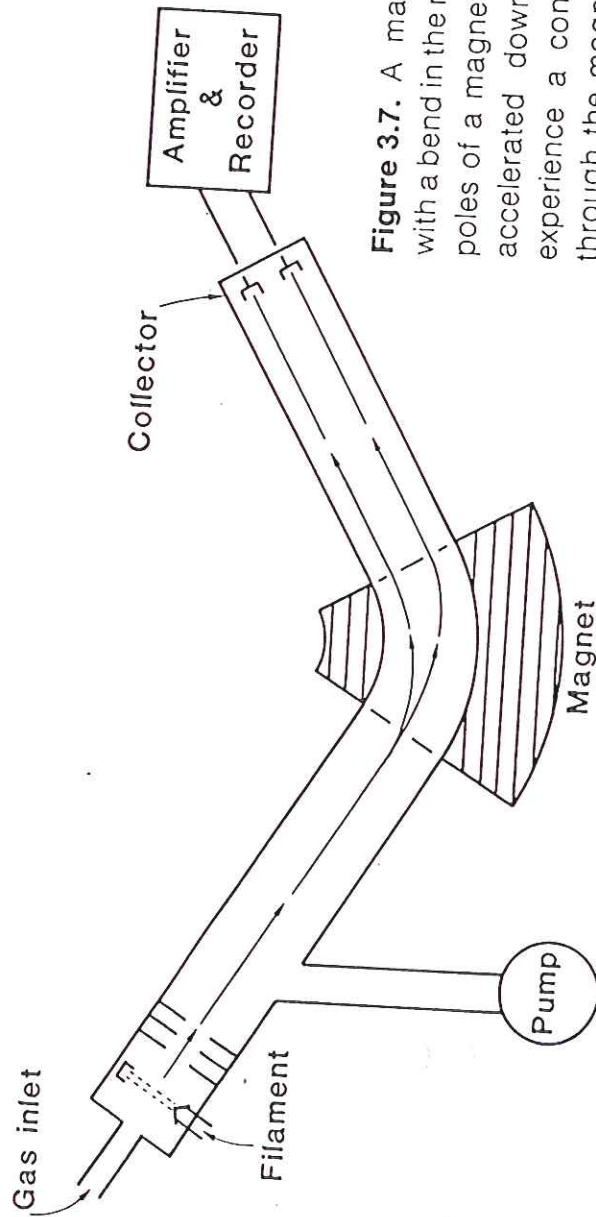


Figure 3.7. A mass spectrometer is a tube with a bend in the middle between the opposite poles of a magnet. Atomic or molecular ions accelerated down tube by a voltage drop experience a constant force while passing through the magnet and are thus separated into different beams according to mass.

TABLE 3.1
*Principal Parent and Daughter Isotopes Used to
 Determine the Ages of Rocks and Minerals*

Parent isotope (radioactive)	Daughter isotope (stable)	Half-life (Ma)	Decay constant (yr^{-1})
^{40}K	$^{40}\text{Ar}^a$	1,250	5.81×10^{-11}
^{87}Rb	^{87}Sr	48,800	1.42×10^{-11}
^{147}Sm	^{143}Nd	106,000	6.54×10^{-12}
^{176}Lu	^{176}Hf	35,900	1.93×10^{-11}
^{187}Re	^{187}Os	43,000	1.612×10^{-11}
^{232}Th	^{208}Pb	14,000	4.948×10^{-11}
^{235}U	^{207}Pb	704	9.8485×10^{-10}
^{238}U	^{206}Pb	4,470	1.55125×10^{-10}

^a ^{40}K also decays to ^{40}Ca , for which the decay constant is $4.962 \times 10^{-10} \text{ yr}^{-1}$, but that decay is not used for dating. The half-life is for the parent isotope and so includes both decays.

Natural Abundances of the Isotopes Used in Radiometric Dating

Isotope	Abundance (%)	Isotope	Abundance (%)
³⁹ K	93.26	³⁶ Ar	0.337
⁴⁰ K	0.0117	³⁸ Ar	0.063
⁴¹ K	6.73	⁴⁰ Ar	99.60
⁸⁵ Rb	72.17	⁸⁴ Sr	0.56
⁸⁷ Rb	27.83	⁸⁶ Sr	9.87
		⁸⁷ Sr	7.04
		⁸⁸ Sr	82.53
¹⁴⁴ Sm	3.0	¹⁴² Nd	27.3
¹⁴⁷ Sm	14.9	¹⁴³ Nd	12.3
¹⁴⁸ Sm	11.2	¹⁴⁴ Nd	23.8
¹⁴⁹ Sm	13.8	¹⁴⁵ Nd	8.3
¹⁵⁰ Sm	7.4	¹⁴⁶ Nd	17.1
¹⁵² Sm	26.8	¹⁴⁸ Nd	5.7
¹⁵⁴ Sm	22.9	¹⁵⁰ Nd	5.6
¹⁷⁵ Lu	97.4	¹⁷⁴ Hf	0.17
¹⁷⁶ Lu	2.6	¹⁷⁶ Hf	5.2
		¹⁷⁷ Hf	18.5
		¹⁷⁸ Hf	27.2
		¹⁷⁹ Hf	13.8
		¹⁸⁰ Hf	35.1
¹⁸⁵ Re	37.40	¹⁸⁴ Os	0.02
¹⁸⁷ Re	62.6	¹⁸⁶ Os	1.6
		¹⁸⁷ Os	1.6
		¹⁸⁸ Os	13.3
		¹⁸⁹ Os	16.1
		¹⁹⁰ Os	26.4
		¹⁹² Os	41.0
²³² Th	100.0	²⁰⁴ Pb	1.4
		²⁰⁶ Pb	25.2
²³⁴ U	0.0057	²⁰⁷ Pb	21.7
²³⁵ U	0.72	²⁰⁸ Pb	51.7
²³⁸ U	99.27		

NOTE: Abundances are for the Earth's crust except for argon, which is for the atmosphere. The isotopic abundances for those elements that include a daughter isotope vary because of decay of the corresponding parent isotope. The isotope pairs used in radiometric dating are indicated by arrows.

SOURCES: Lederer, Holland, and Perlman, 1967; Faure, 1986.

As noted, Rb-Sr particularly useful for very old rocks

Age of the Earth — could try to find oldest rock

Contest among geochronologists — need to find rocks that have not been re-heated since — like track records

Show map Fig. 16.6 of ages of continental crust — basement rock beneath veneer of sediments

Archean — > 2500 my = 2.5 billion
in black

~~the~~ Such old terranes are geologically complex — oldest rocks often fragmentary inclusions of igneous rocks or metasedimentary rocks

~~Three examples~~ • ~~Greenland, Labrador~~
~~connected at that~~
~~time~~
• ~~South Africa~~
• ~~Western Australia~~

Three examples :

- South Africa — map & ³ examples
 $^{40}\text{Ar} / ^{39}\text{Ar}$ & ~~Rb-Sr~~
 3.5 b.y. Rb-Sr isochron
 Nd-Sm isochron
- Western Australia — map & Nd-Sm example
- Greenland & Labrador
 map & 3 examples
 including 3.75 b.y. old
 Nd-Sm date on Isua rocks
 at one time the record
 holder — now somewhere
 in Canada

Phanerozoic < 570 my
 Proterozoic 570 - 2500 my
 Archean > 2500 My = 2.5 By

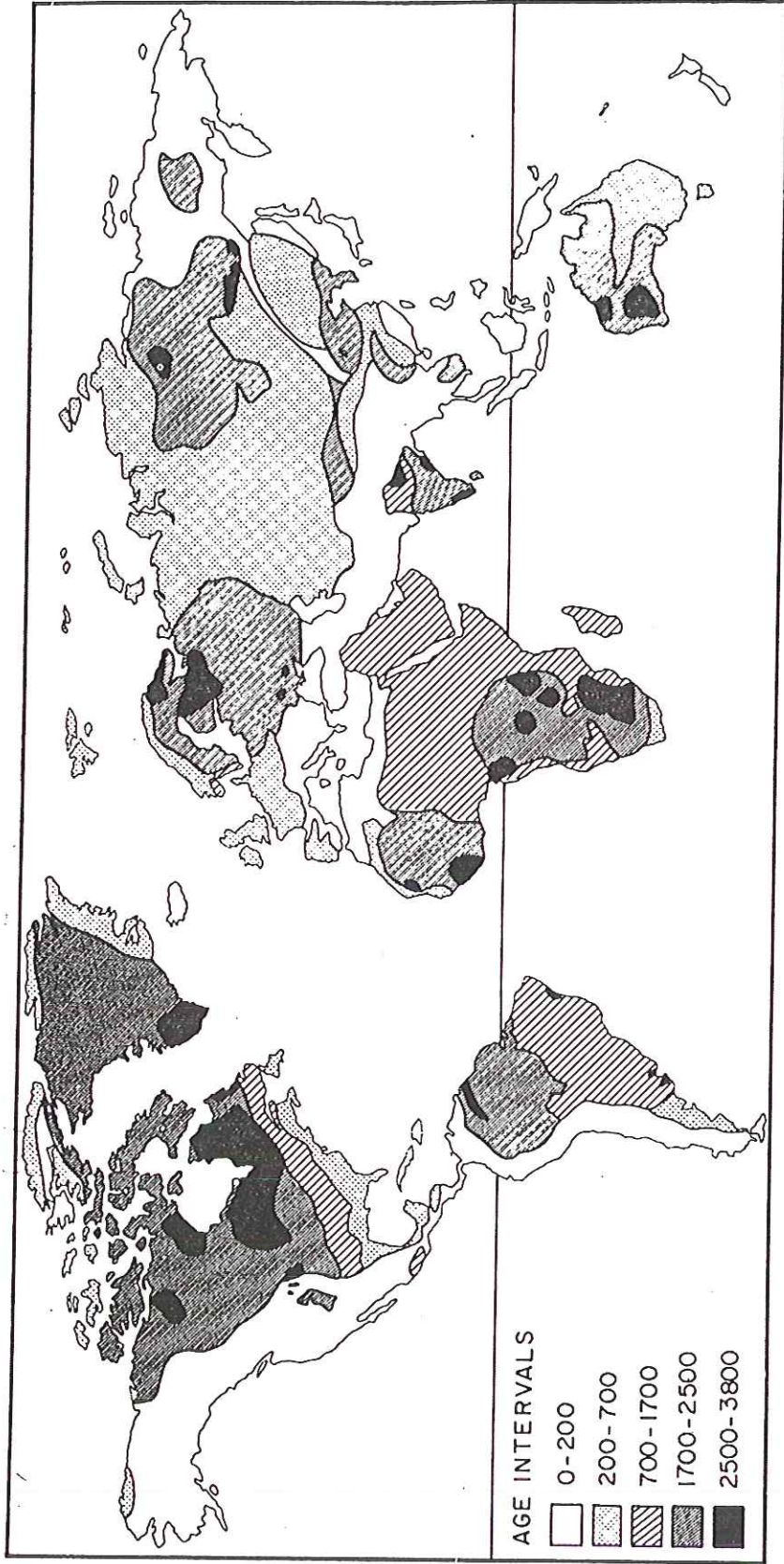


Figure 16.6. Map of Earth showing approximate ages of continental material, in millions of years. Reproduced from Broecker (1985) by permission of Eldigio Press.

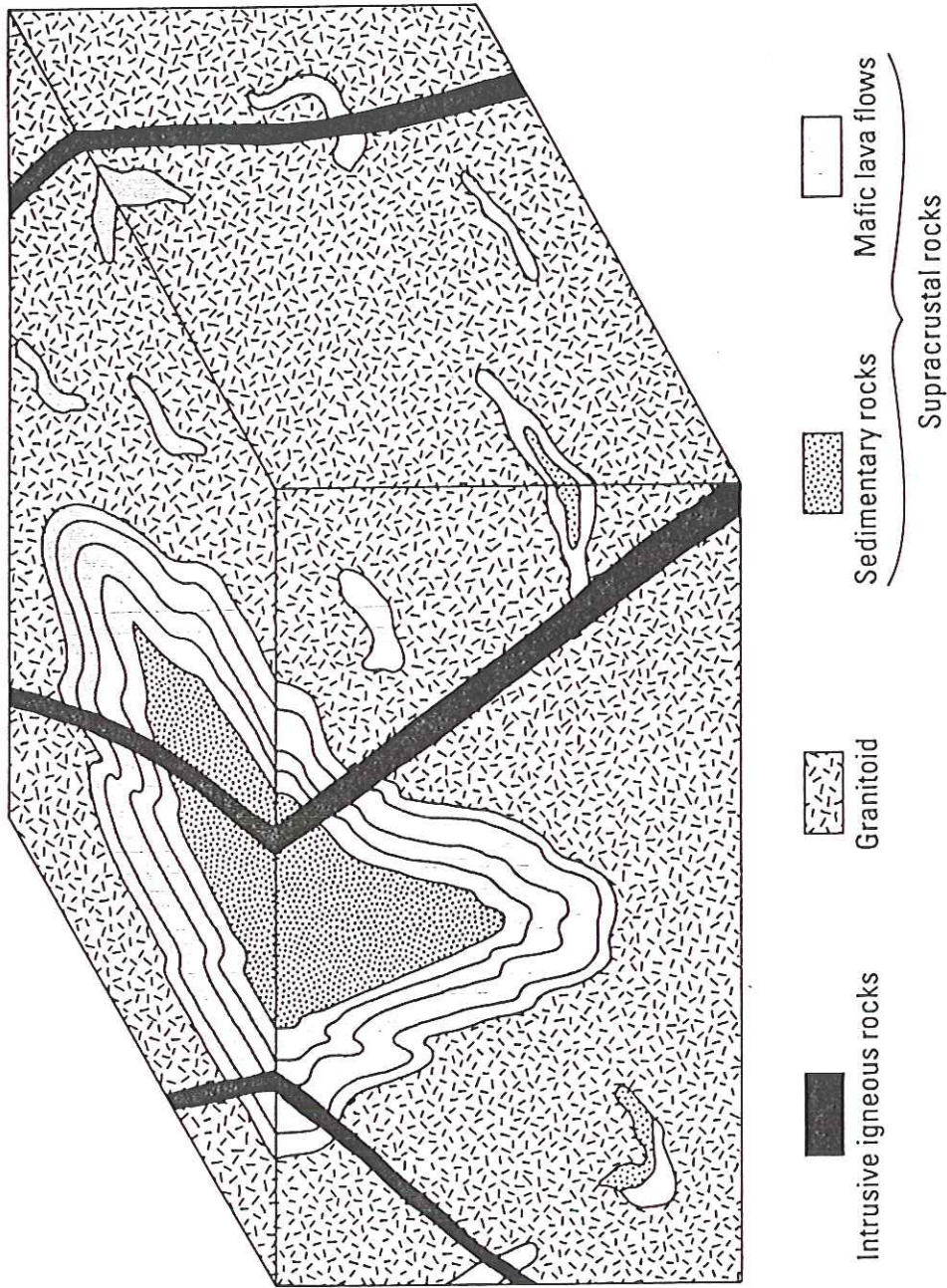


Fig. 4.3. Typical sequence in Archean terranes where the Earth's oldest rocks are found. Commonly, the oldest rocks, which occur as fragmented inclusions within the gneiss, are remnants of lava flows and of sedimentary rocks derived from still older rocks of which there is now no trace.

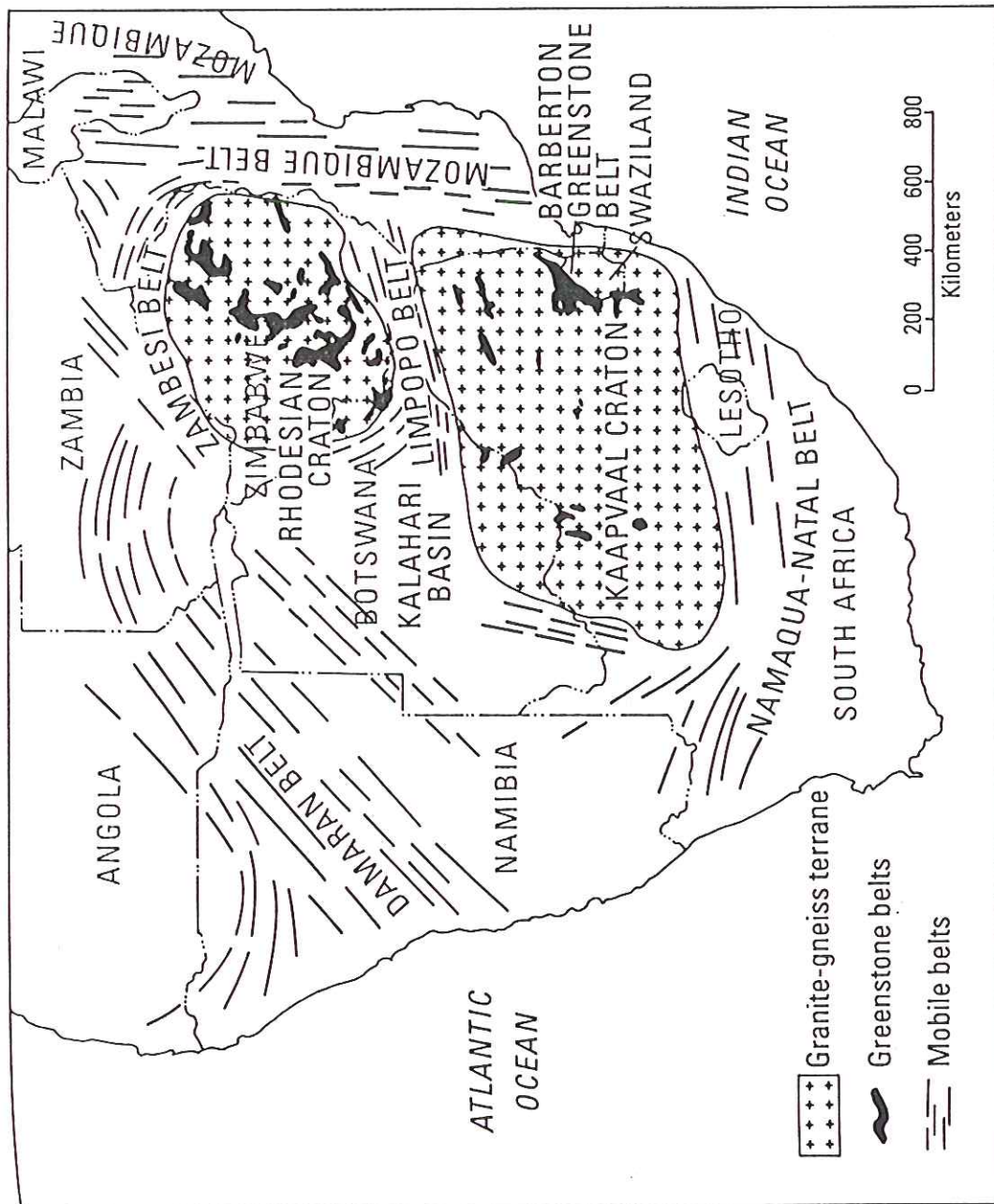


Fig. 4.26. The Archean cratons of southern Africa. The boundaries of the Kaapvaal and Rhodesian cratons are uncertain because large areas are concealed beneath younger rocks. (After R. Mason, 1973.)

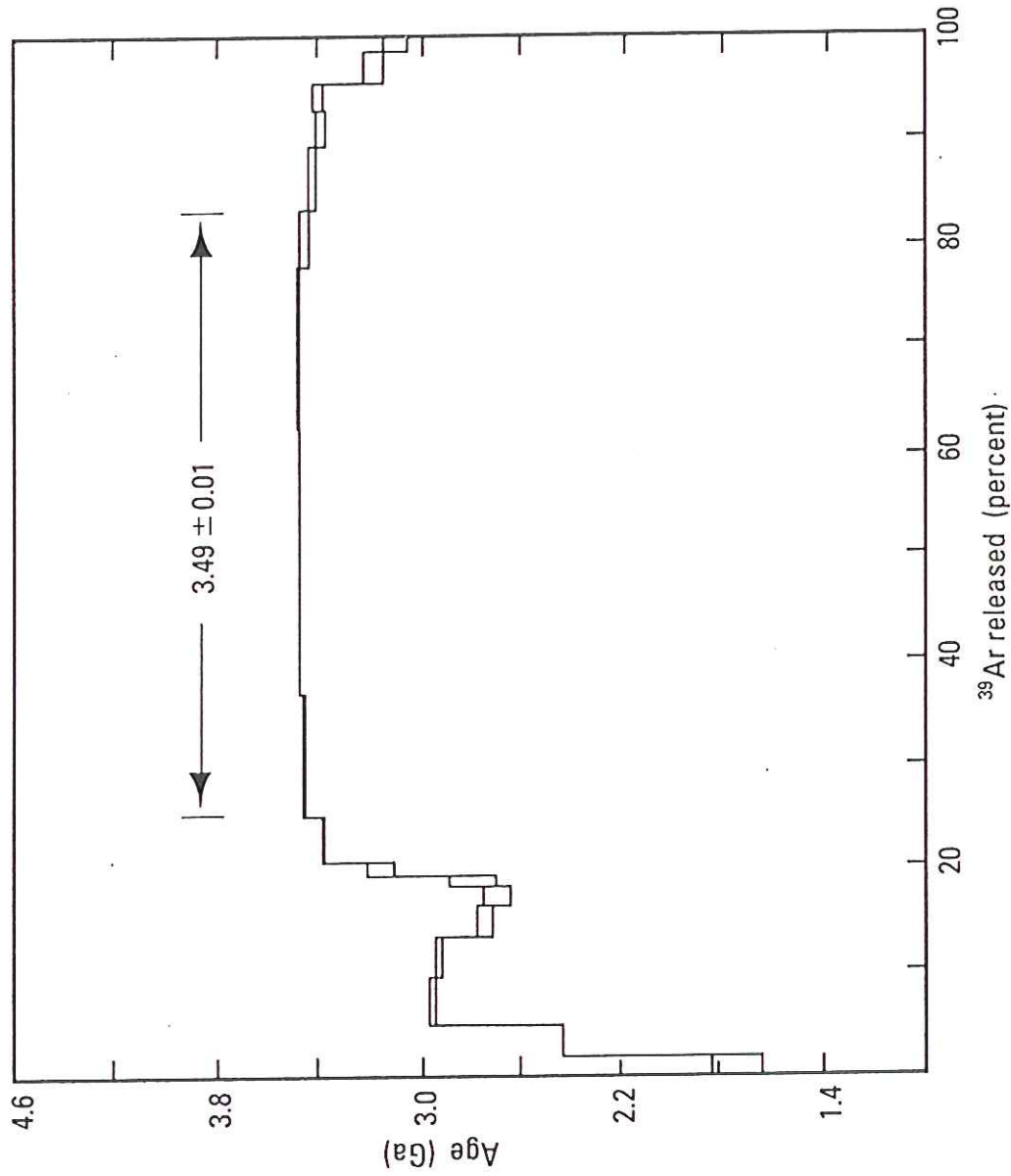


Fig. 4.29. $^{40}\text{Ar}/^{39}\text{Ar}$ Ar age spectrum for a sample of komatiite from the Komati Formation, Onverwacht Group, Barberton Mountain Land, southern Africa. The vertical thickness of the boxes indicates the standard deviation of the value for each gas increment. Error in the plateau age indicates two standard deviations. (After M. Lopez Martinez et al., 1984.)

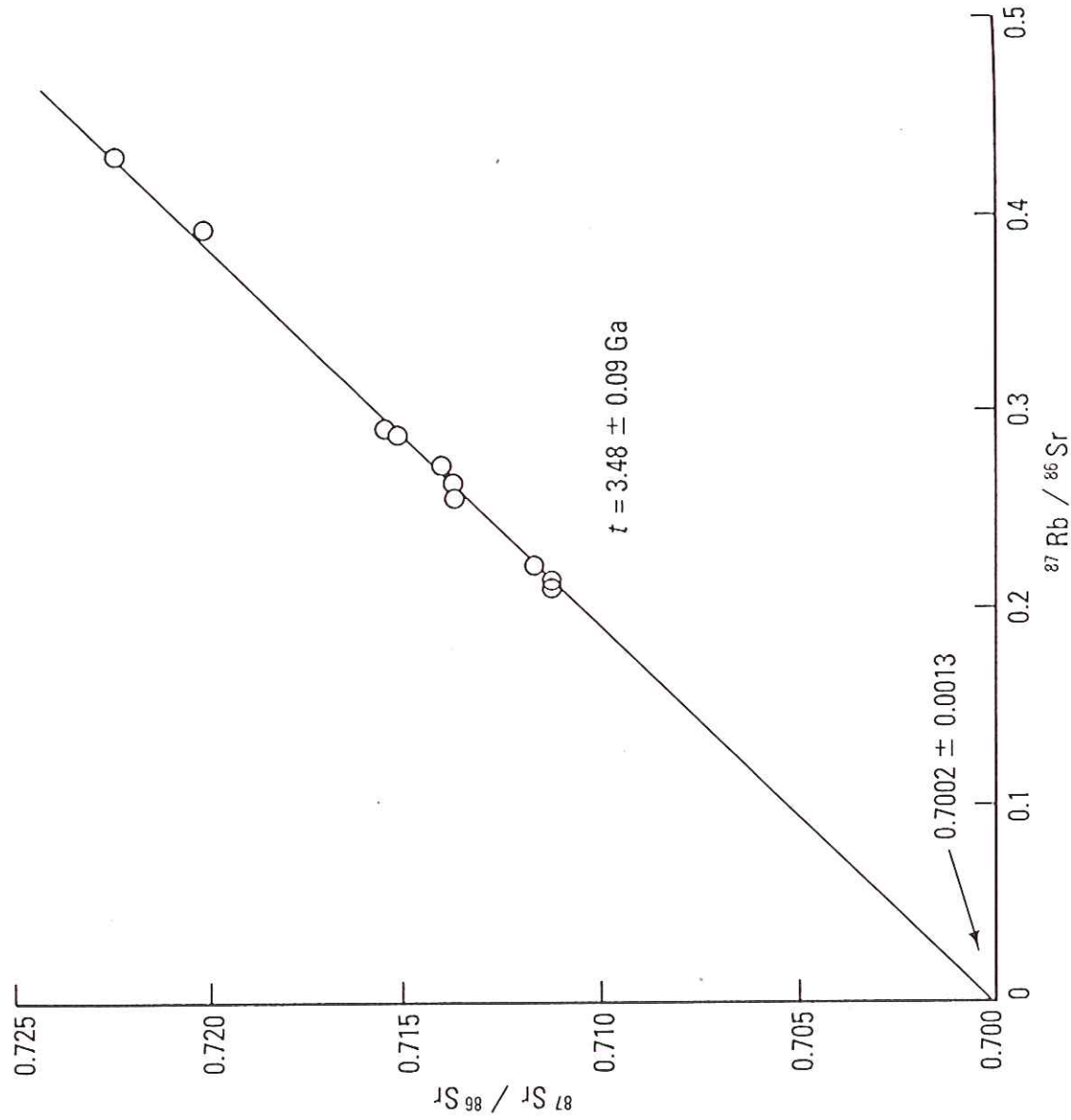


Fig. 4.31. Rb-Sr isochron for whole-rock samples from the Stolzburg Pluton, South Africa. (After Barton et al., 1983b.)

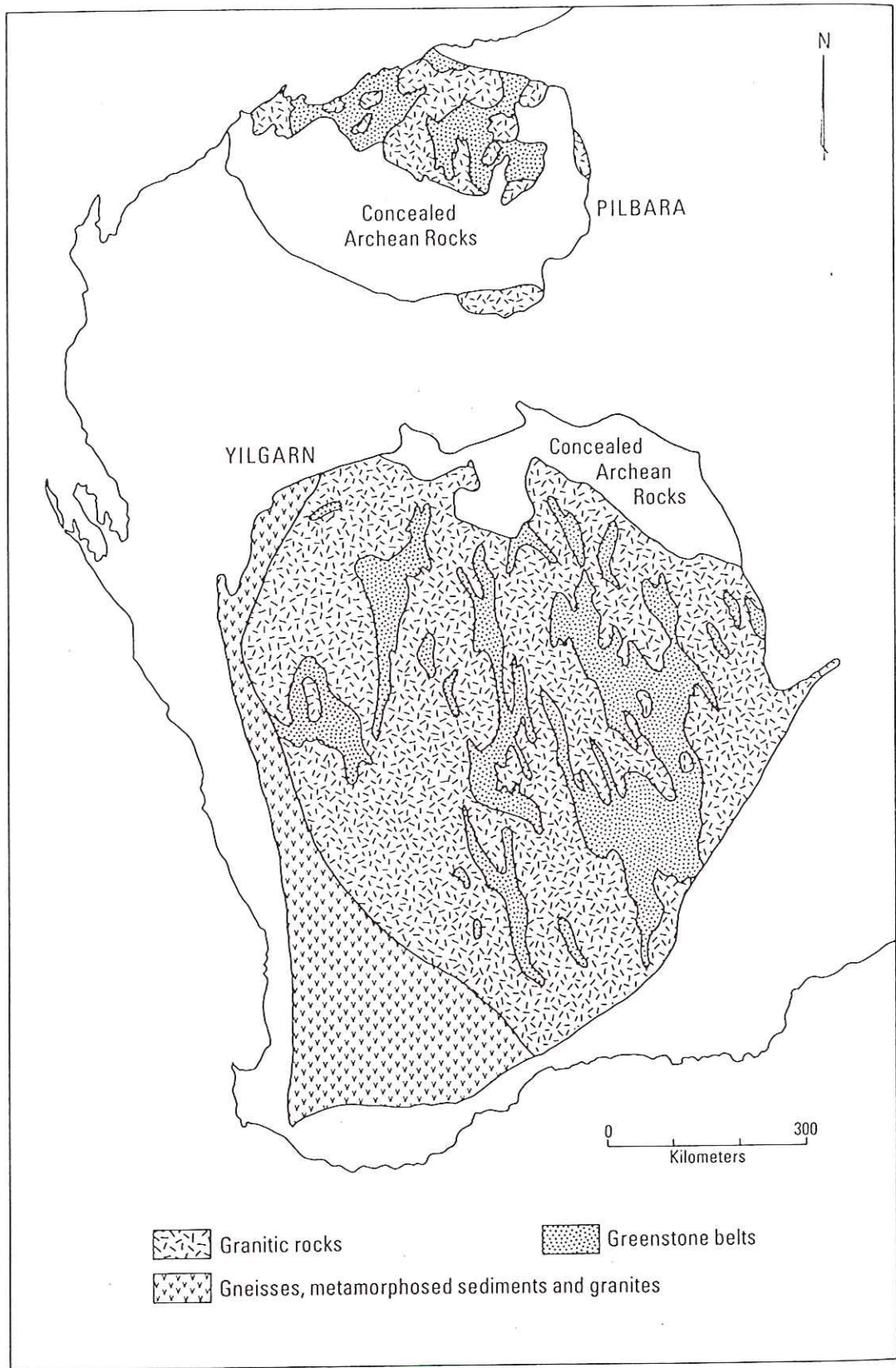


Fig. 4.21. Principal rock types of the Archean Pilbara and Yilgarn blocks, Western Australia. (After Rutland, 1981.)

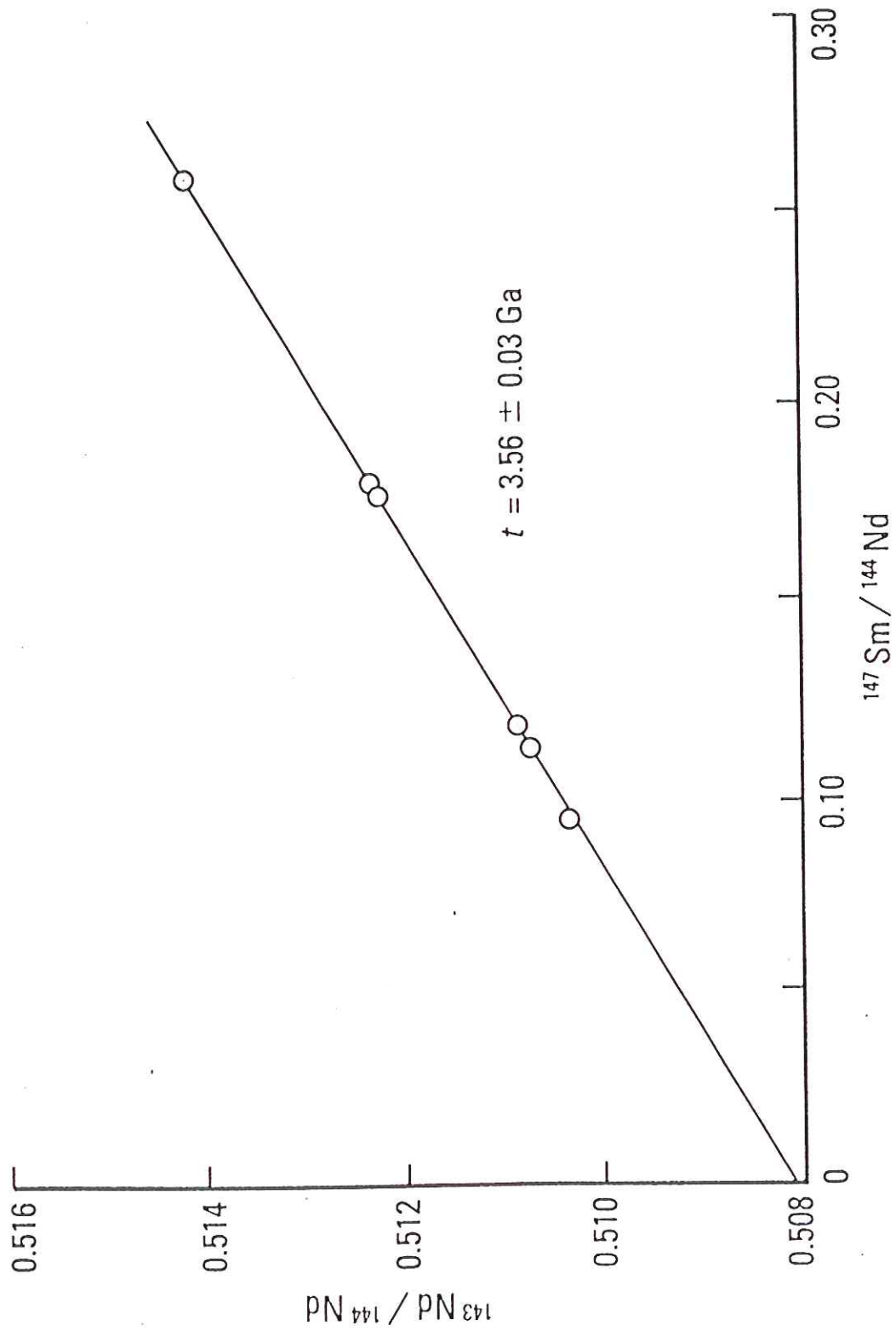


Fig. 4.23. Sm-Nd isochron diagram for samples of volcanic rocks from the North Star Basalt, the lowest formation in the Pilbara Supergroup, Pilbara Block, Western Australia. (After Hamilton et al., 1981.)

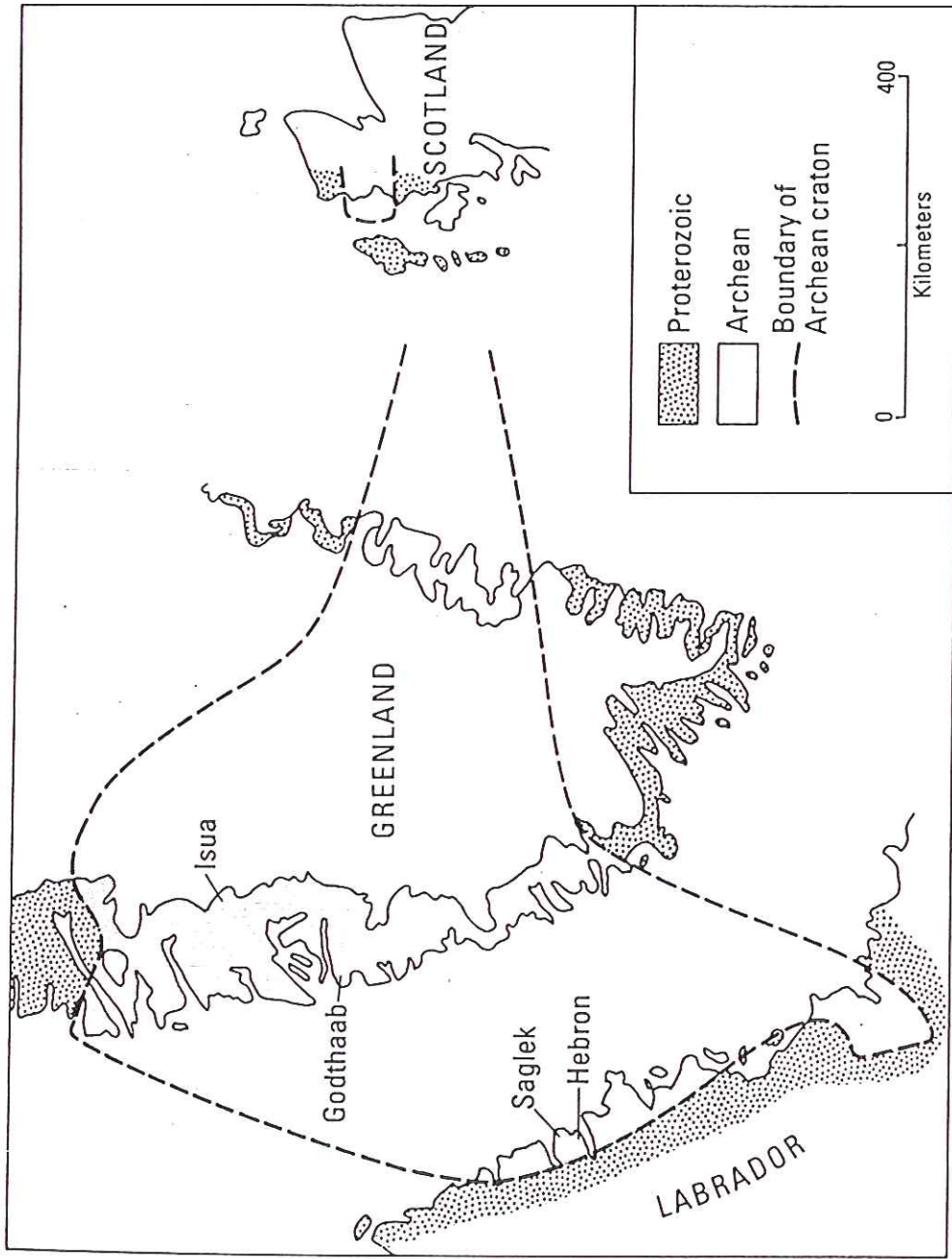


Fig. 4.6. The North Atlantic craton includes parts of Greenland, Labrador, Scotland, and Norway (not shown). The land masses, joined prior to about 200 Ma, are shown closer together than they are now. Precambrian rocks do not occur in the oceans that now divide the fragments of the craton. (After Bridgwater, Watson, and Windley, 1973.)

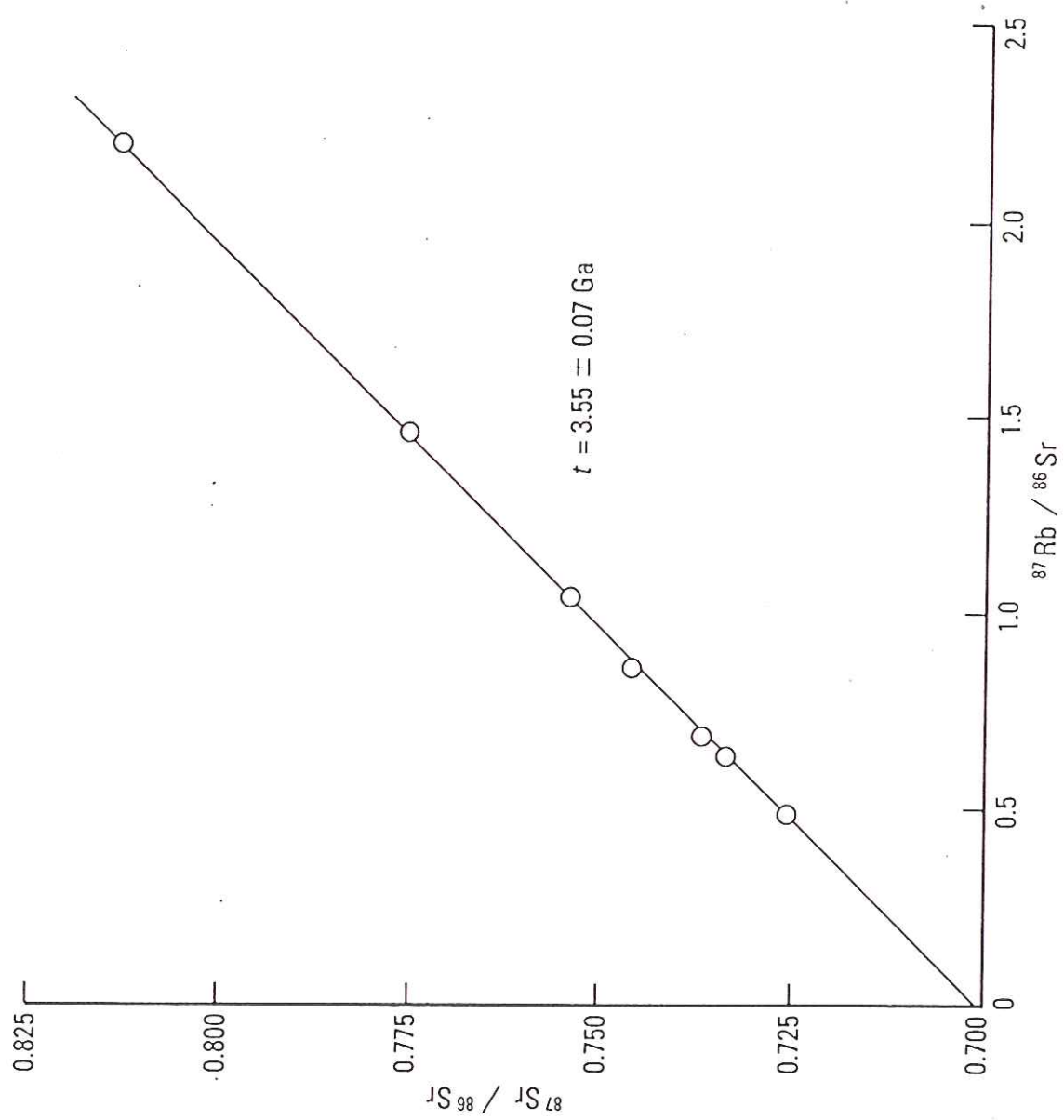


Fig. 4.16. Rb-Sr isochron diagram for samples of the Uivak gneisses near Saglek, eastern Labrador. (After Hurst et al., 1975.)

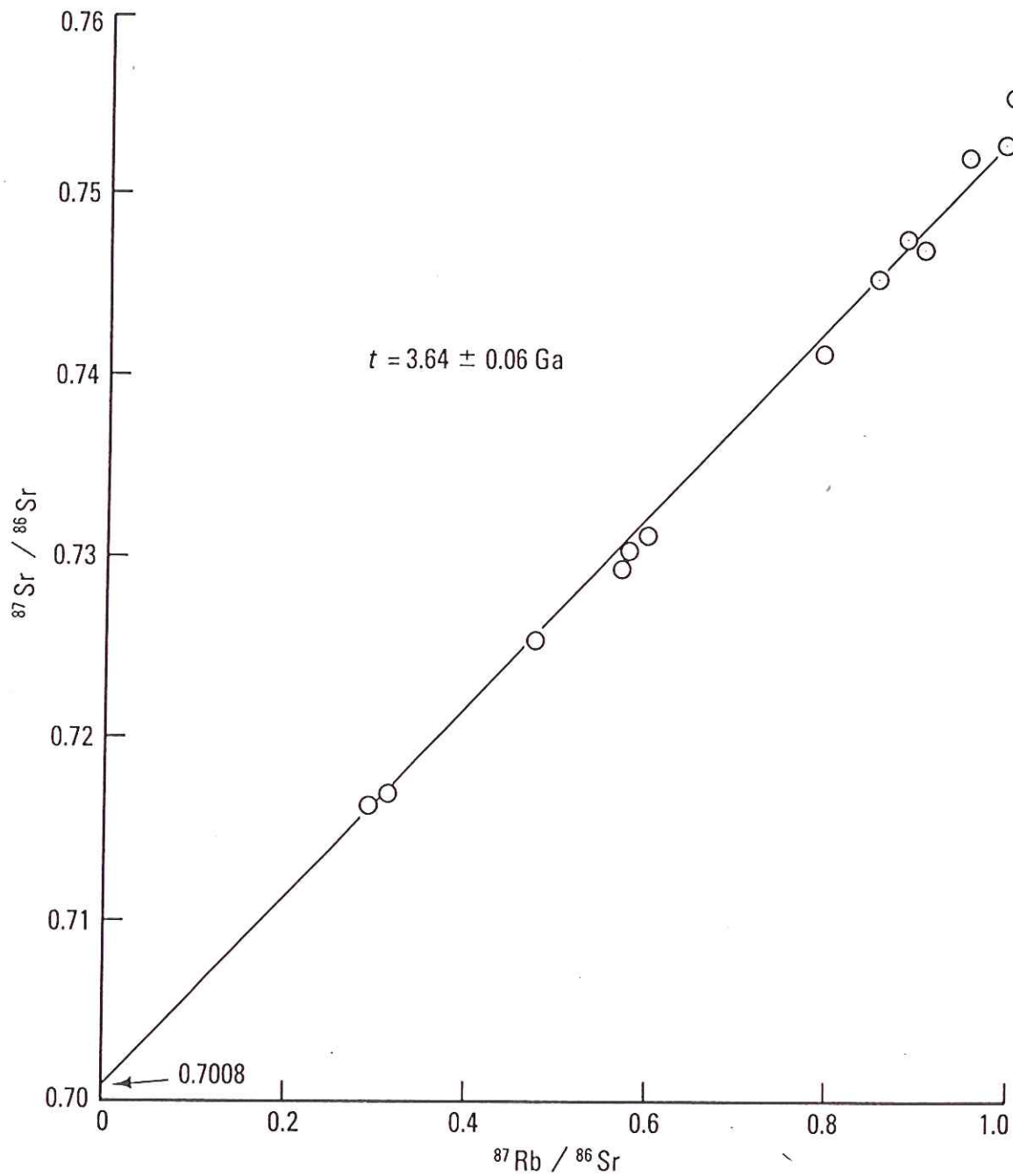


Fig. 4.14. Rb-Sr isochron diagram for samples of the Amitsoq gneisses at Isua. (After Moorbath et al., 1977a.)

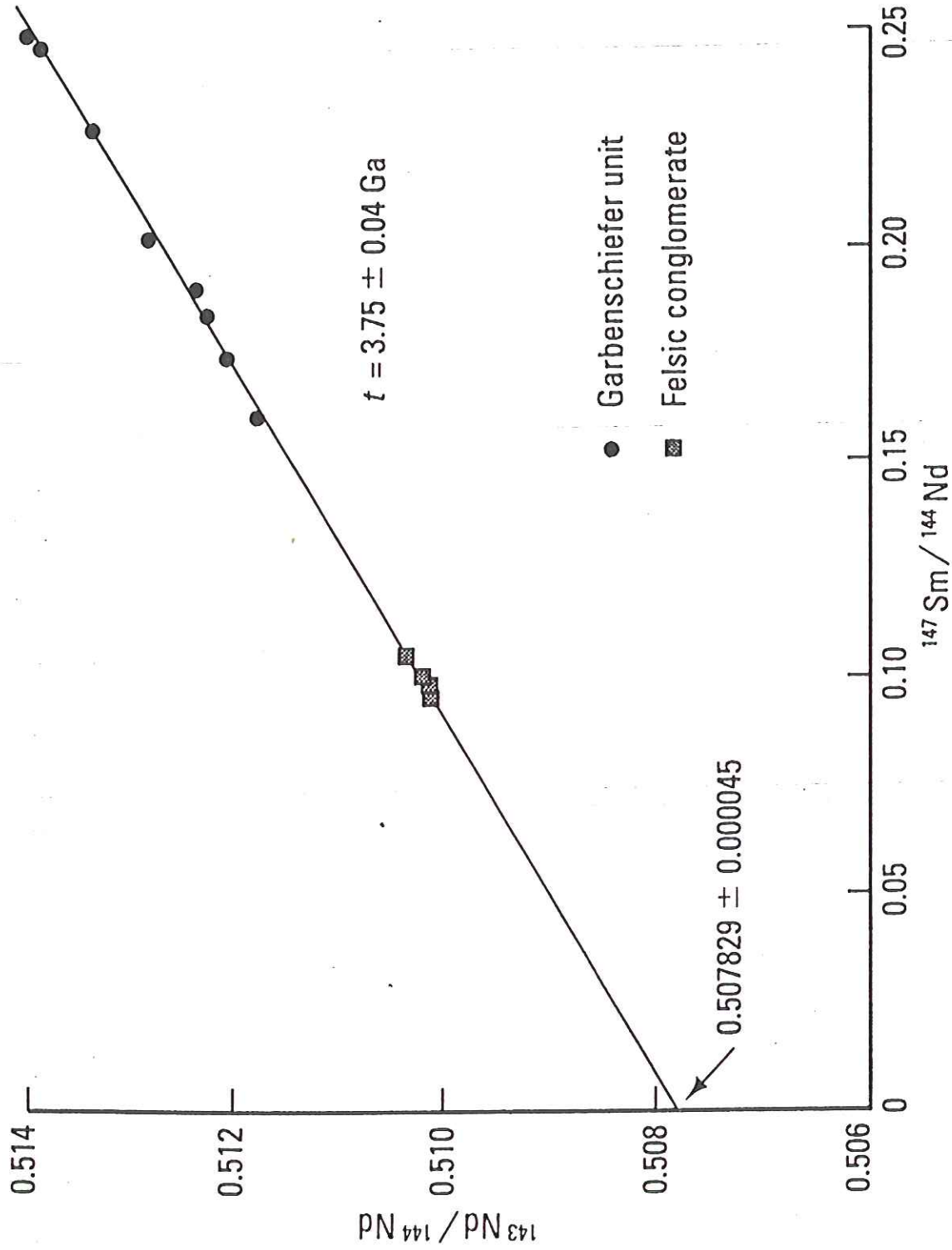
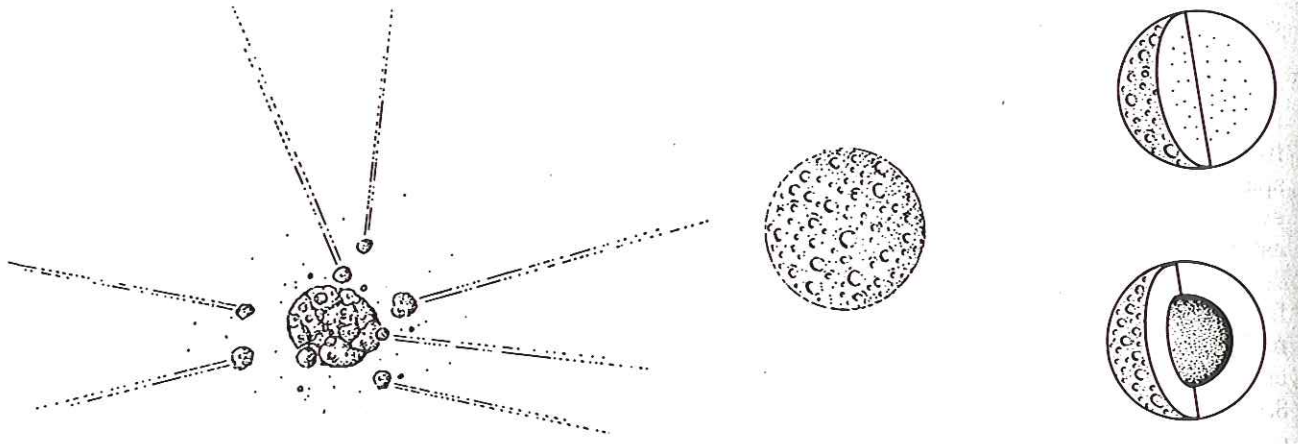
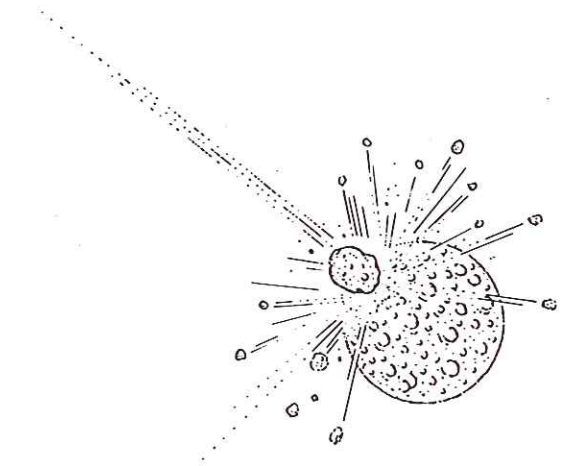


Fig. 4.11. Sm-Nd isochron diagram for samples from the garbenschiefer unit and the felsic conglomerate of the Isua supracrustals. (Data from Hamilton et al., 1978, 1983.)

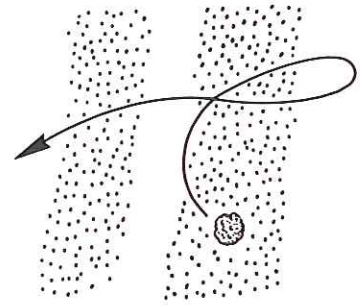


(A) Accretion of a planetesimal 4.6 billion years ago.

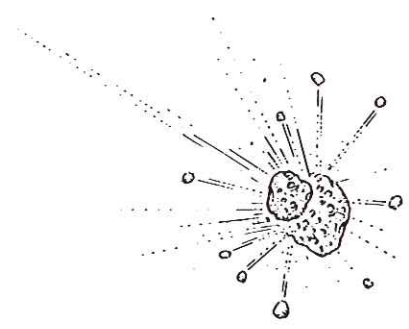
(B) Differentiation of interior to form iron core or metamorphosed rock 4.4 to 4.6 billion years ago.



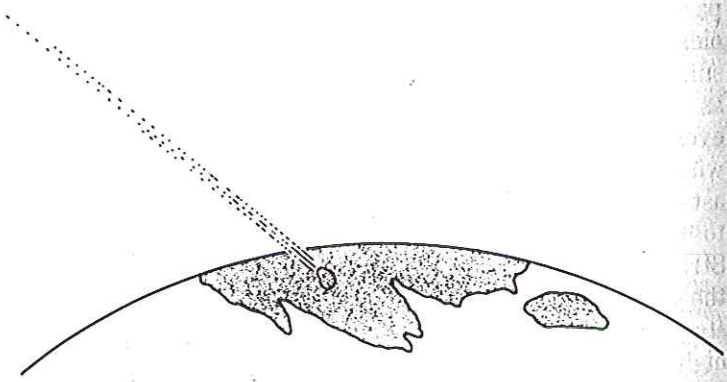
(C) Collision and fragmentation 1 to 0.1 billion years ago.



(D) Deflection into inner solar system 1.0 to 0.1 billion years ago.



(E) Further collisions less than 0.1 billion years ago.



(F) Impact on Earth today.

Figure 3.15
The evolution of asteroids as meteorite parent bodies is summarized in this diagram. The events that led to the delivery of fragments of the asteroids to Earth and other inner planets are emphasized.

To get the true age of the Earth we turn to meteorites — particularly the carbonaceous chondrites which we believe to be fossils of the primitive solar nebula

All types of age dating methods have been applied to meteorites with the same result:

Broecker Fig. 4-2

- initial $^{87}\text{Sr}/^{86}\text{Sr}$ in solar nebula = 0.7

$$\frac{700 \text{ } ^{87}\text{Sr}}{1000 \text{ } ^{86}\text{Sr}} \text{ atoms}$$

- age of solar system 4.5-4.6 b.y.

treats the whole solar nebula as one big rock — 38 meteorites or "mineral"

Fig. C.11 — data for 38 different meteorites

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}} \right)_0 = 0.69885 \pm 0.00010$$

$$t = 4.498 \pm 0.015 \text{ b.y.}$$

very accurate!

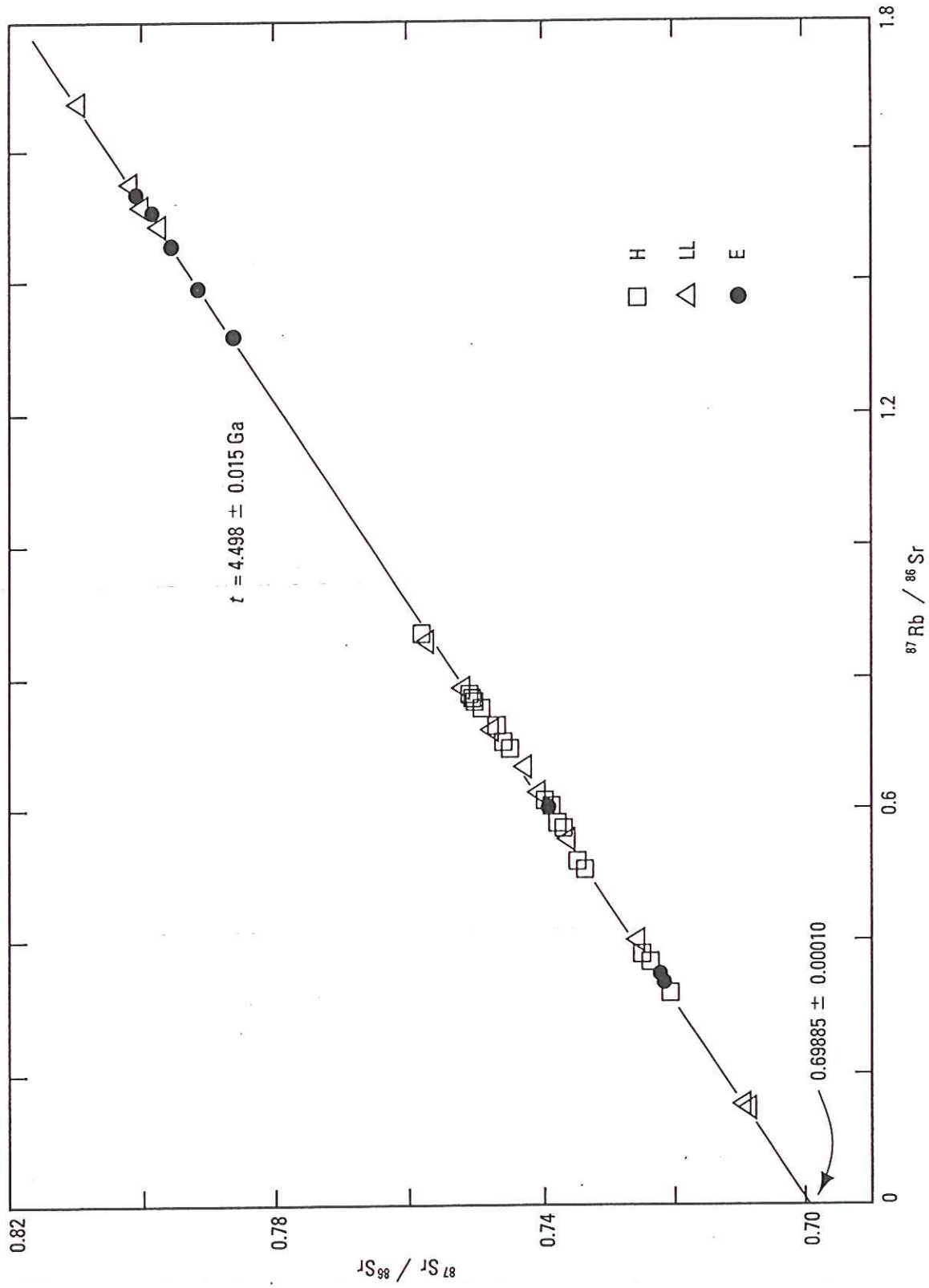


Fig. 6.11. Rb-Sr whole-rock isochron for 38 undisturbed H, LL, and E chondrites. (After Minster, Birck, and Allègre, 1982.)

This figure is a plot of the ratio of ^{87}Sr to ^{86}Sr versus the ratio of ^{87}Rb to ^{86}Sr for 38 undisturbed H, LL, and E chondrites. The data points are categorized by H (open squares), LL (open triangles), and E (filled circles). A linear fit is shown with a slope of $t = 4.498 \pm 0.015 \text{ Ga}$ and an intercept of 0.69885 ± 0.00010 .

shows same data (Minster, Birck & Allegre)

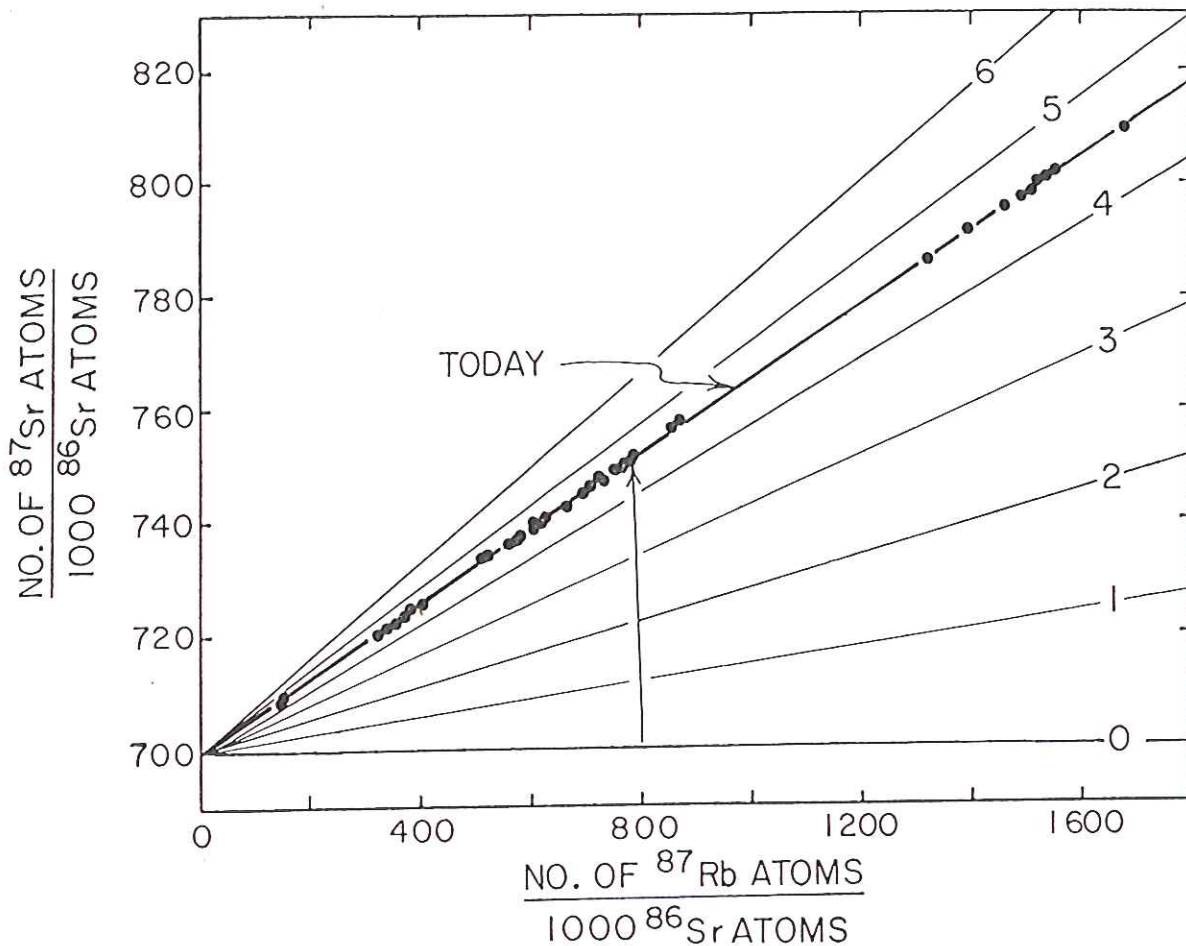


Figure 4-2. Evolution of strontium isotope composition in minerals of differing rubidium contents: The light lines on this diagram show the evolution with time (billions of years) of the ^{87}Sr and ^{87}Rb in meteorites. The measurements on mineral grains separated from chondritic meteorites tell us two things. First, they tell us that there were 700 ^{87}Sr atoms for each 1000 ^{86}Sr atoms in the strontium present in the solar nebula. Second, they tell us that these meteorites formed very close to 4.56 billion years ago. The former is derived from the intercept of the straight line that passes through the measured values. The latter is derived from the slope of the line passing through these points. Each grain followed a time trend parallel to that for the arrow shown on the diagram. At the time the solar system formed all the grains had compositions falling along the line marked zero, i.e., they had a range of ^{87}Rb to ^{86}Sr ratios, but all had 700 ^{87}Sr atoms per 1000 ^{86}Sr atoms. With time each grain increased in ^{87}Sr content (and decreased in ^{87}Rb content). This increase was in proportion to its ^{87}Rb content.

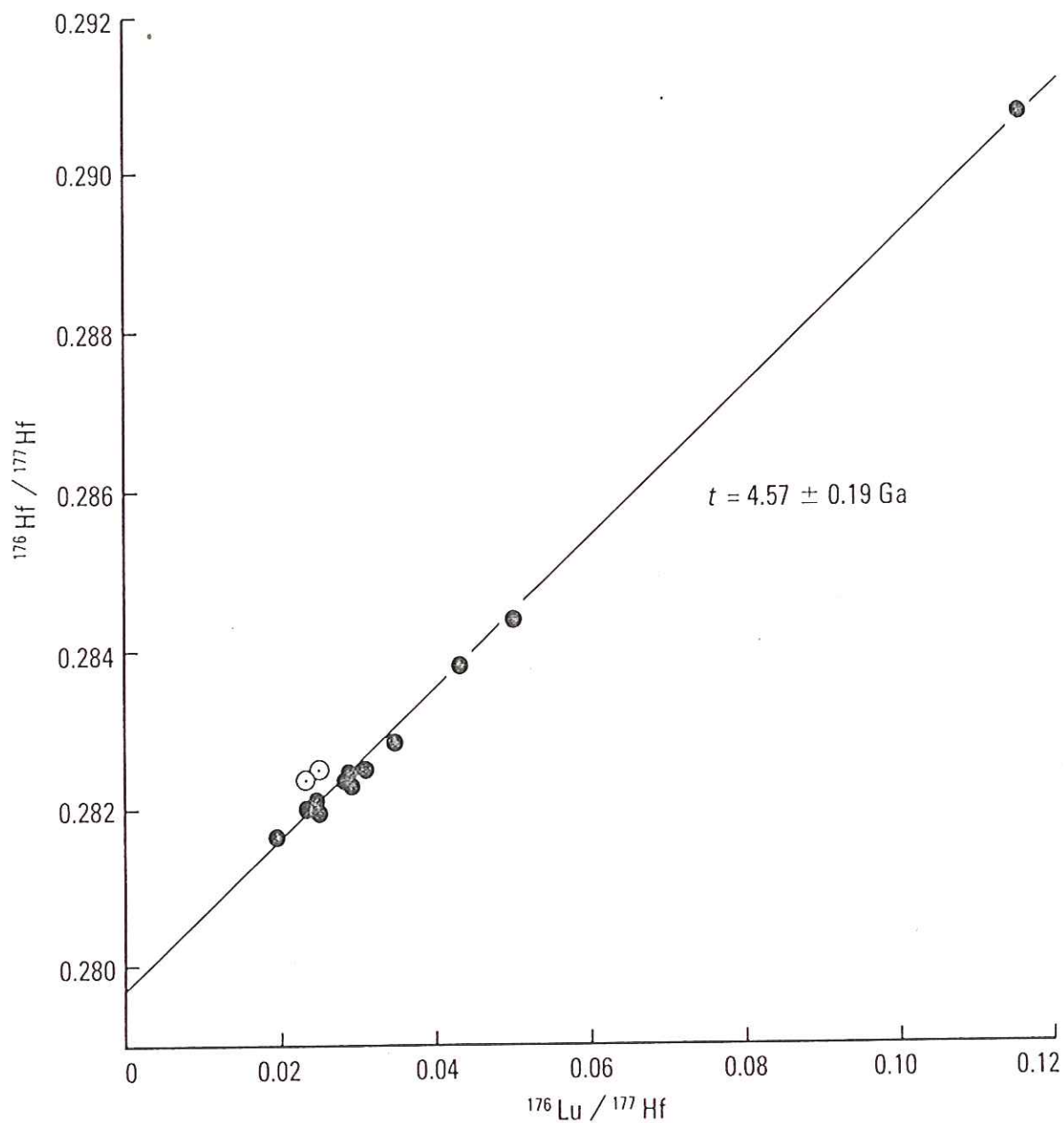


Fig. 6.12. Lu-Hf isochron for 13 eucrites (filled circles). The open circles are data for the antarctic eucrite ALHA77302, which do not fall on the isochron defined by the other eucrites. The age was calculated by using the experimentally determined value for the half-life of ^{176}Lu (Table 3.1). (After Patchett and Tatsumoto, 1980, with additional data from Tatsumoto, Unruh, and Patchett, 1981.)

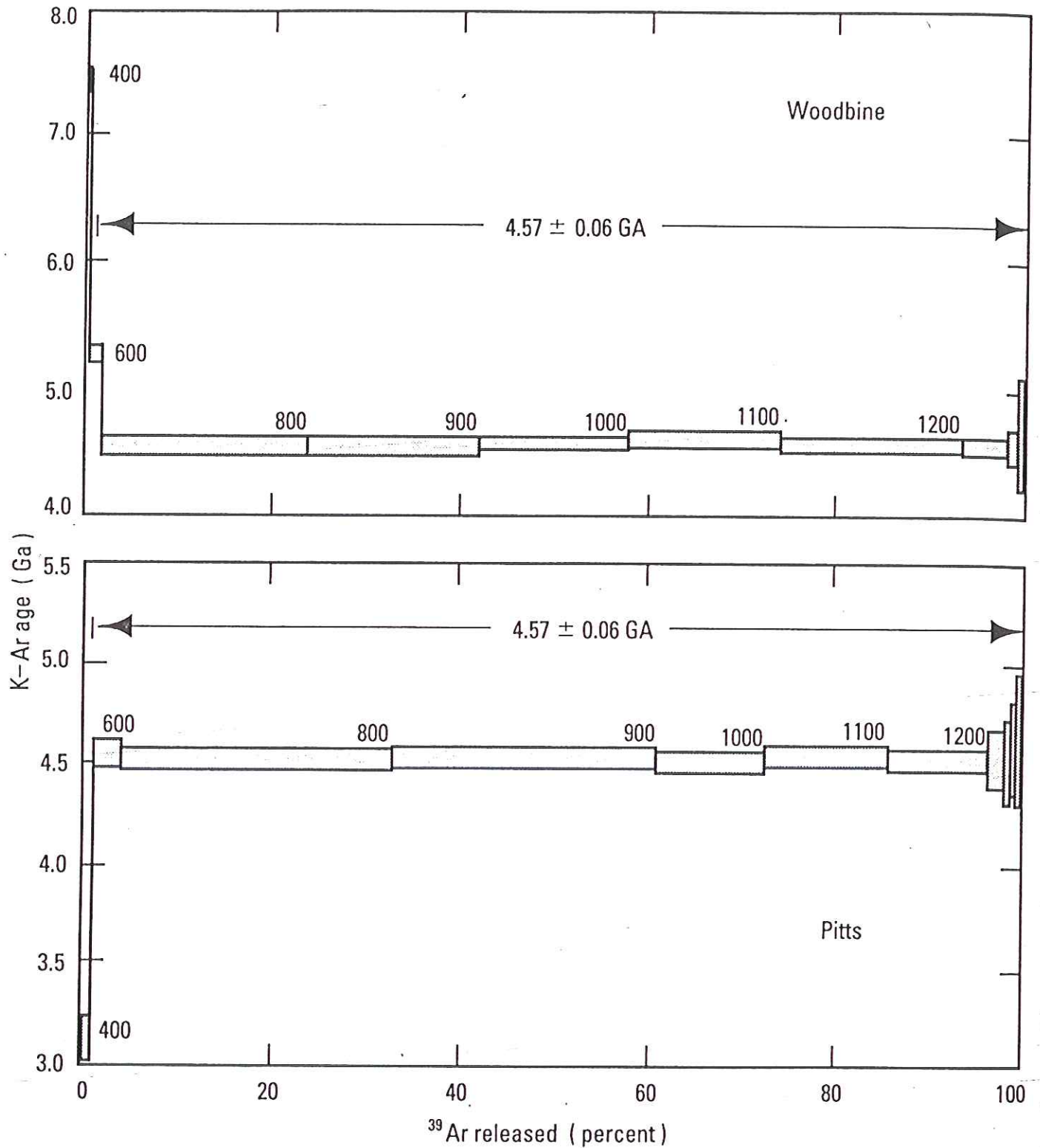


Fig. 6.13. $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra for silicate inclusions in two iron meteorites. (After Niemeyer, 1979.)

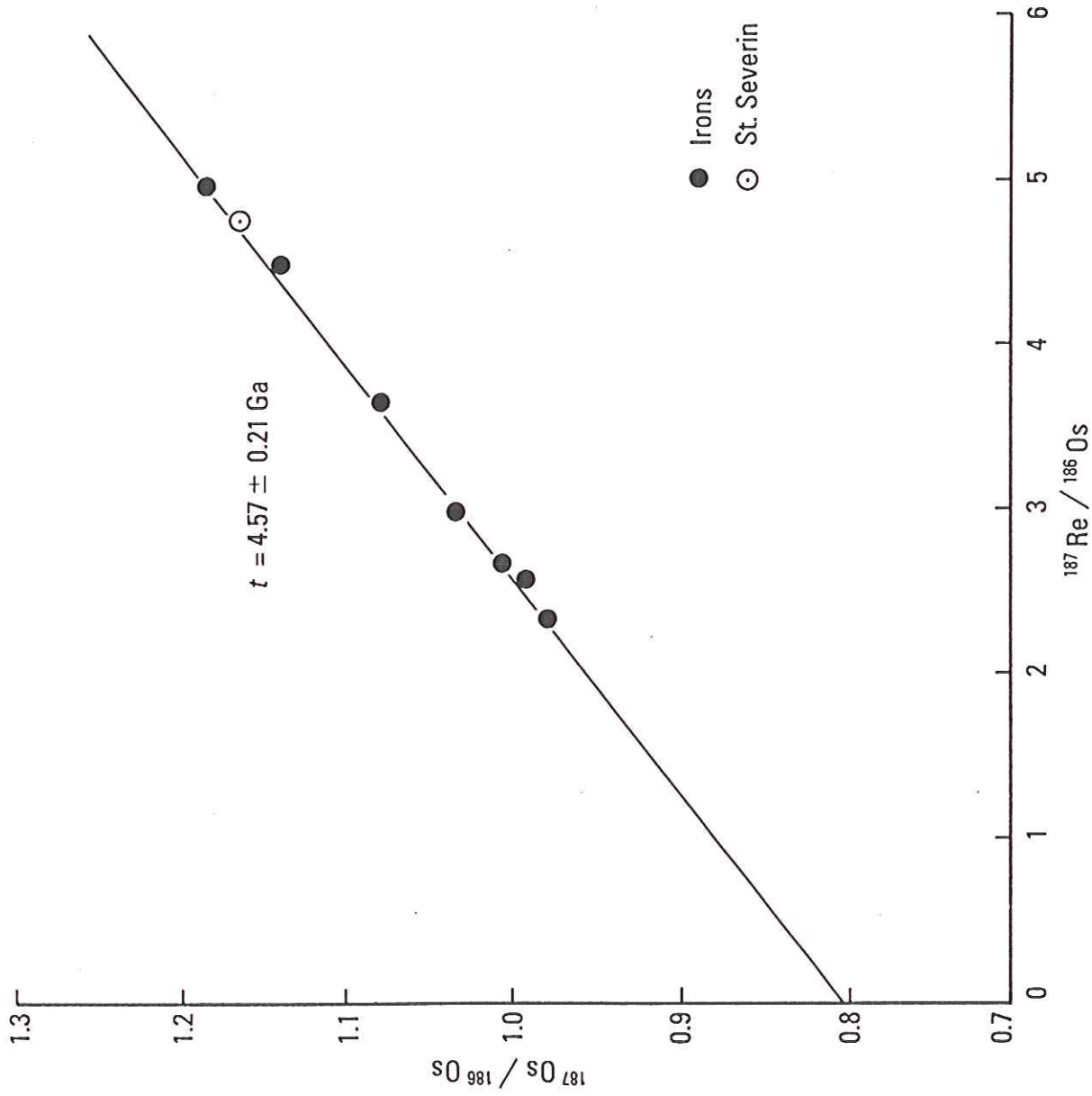


Fig. 6.14. Re-Os isochron for the metal phases of iron meteorites and the St. Severin LL chondrite. The age was calculated by using the experimentally determined value for the half-life of ^{187}Re . (After Luck, Birck, and Allègre, 1980.)

Fig. 6.12 Lu-Hf $t = 4.6 \pm 0.2$ by

Fig. 6.14 Re-Os (metals) in
an iron meteorite
 $t = 4.6 \pm 0.2$ by

Evidence that Fe core
formation in this meteorite
parent body (and by implication
the \oplus) very early

Fig. 6.13 $^{40}\text{Ar}/^{39}\text{Ar}$ age plateaus in
Si inclusions in two Fe
meteorites $t = 4.6 \pm 0.06$ by

Conclude with Fig. 12

Summary of history of solar system

4.5 - 4.6 meteorite ages

3.8 - 3.9 oldest \oplus rocks

15 - 20 big bang

The Moon - of time

Complex history — formed by accretion
or collision of Mars-sized body
with \oplus 4.5 - 4.6 by ago

4.5 - 3.9 continued bombardment

3.9 - 2.5 formation of Mare basins

As on \oplus find rocks of many ages

Resetting by later impact — not by
plate tectonics as on \oplus

H. Schmitt only geologist — knew
what to look for

Sample 72417 dunite — $t = 4.5$ b.y.

I put this in anyway.

Then conclude with Fig 1.2

North

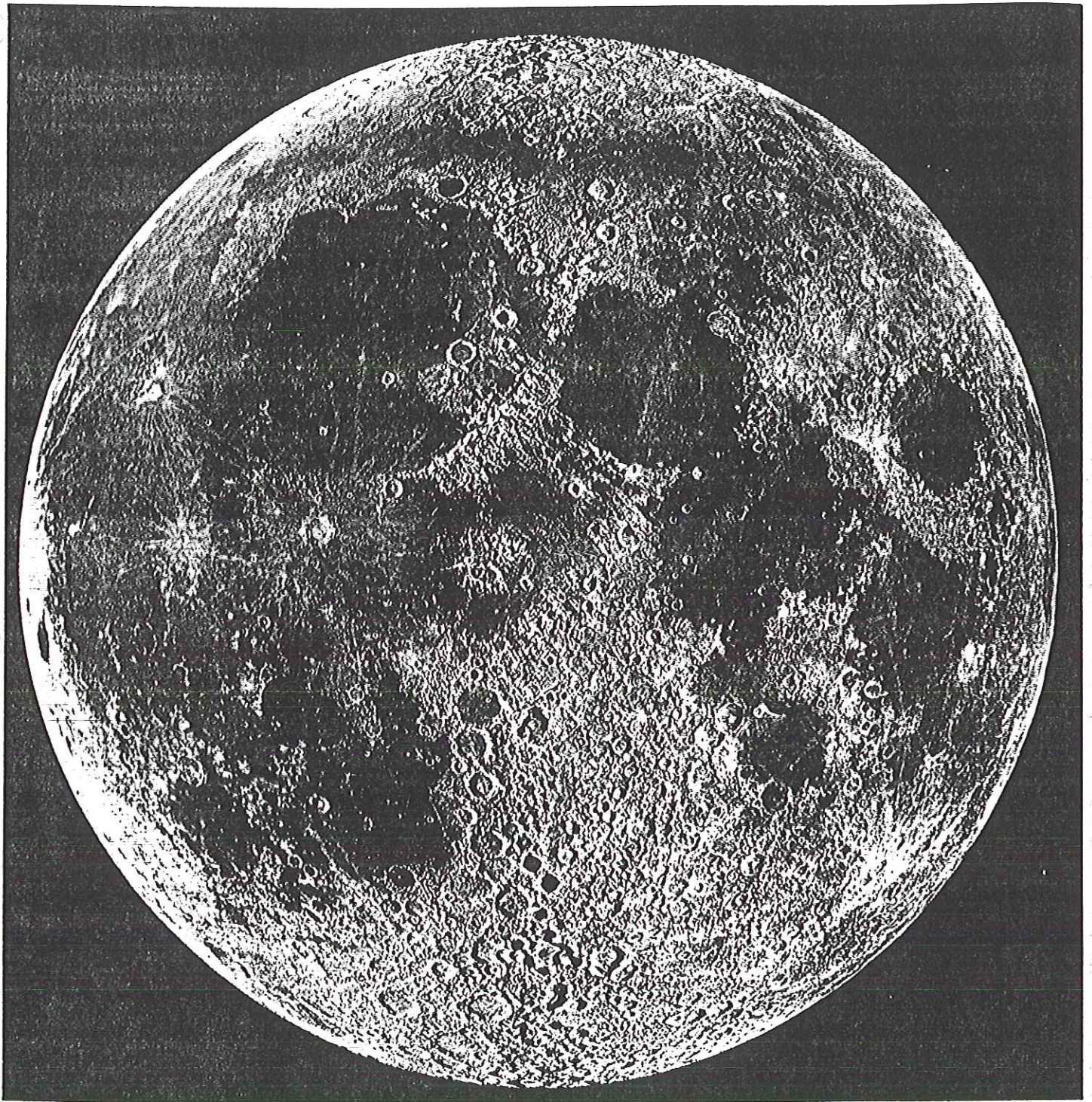


Fig. 5.1. The near side of the Moon, photograph (above) and diagram (opposite), showing some of the principal named features and the locations of the six Apollo (A) and three Luna (L) landing sites from which samples were returned to Earth. The dark smooth areas are maria; the brighter rugged areas are terrae (highlands). (Lick Observatory Photograph L-9, reproduced with permission.)

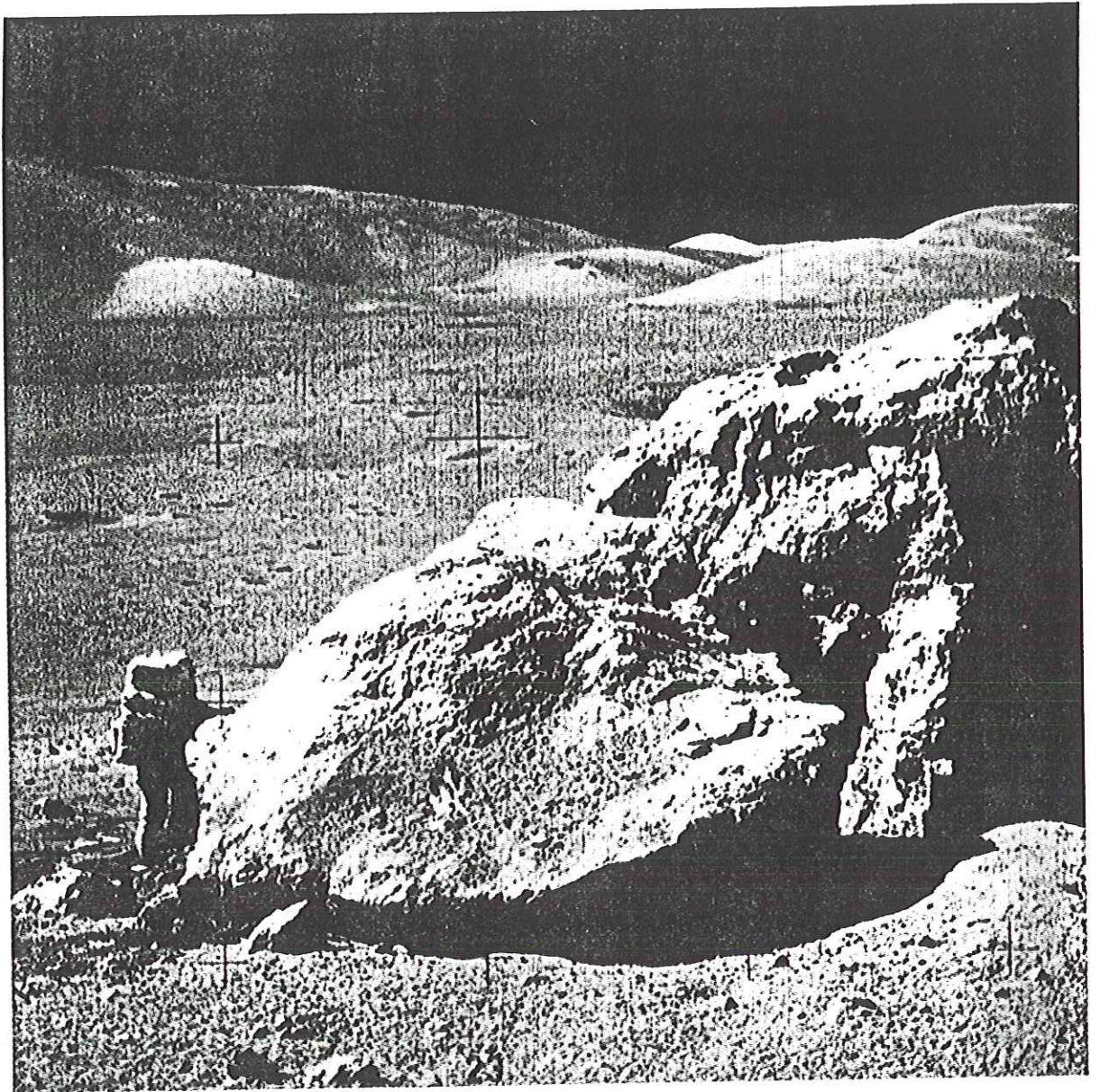


Fig. 5.5. Geologist Harrison H. Schmidt, the only scientist to land on the Moon, examines the large boulder at Station 6 near the Apollo 17 landing site in the Taurus-Littrow Valley. The complex breccia boulder rolled downhill from the adjacent highlands. The ubiquitous lunar regolith, which is composed of comminuted rock, glass, and rock fragments, is the product of repeated meteorite impacts and radiation. None of the rocks visible in the photograph are "in place" where they originally formed. Note the numerous small craters in the regolith. (NASA photograph AS17-140-21497.)

~~AS17-140-21497~~
Apollo 17

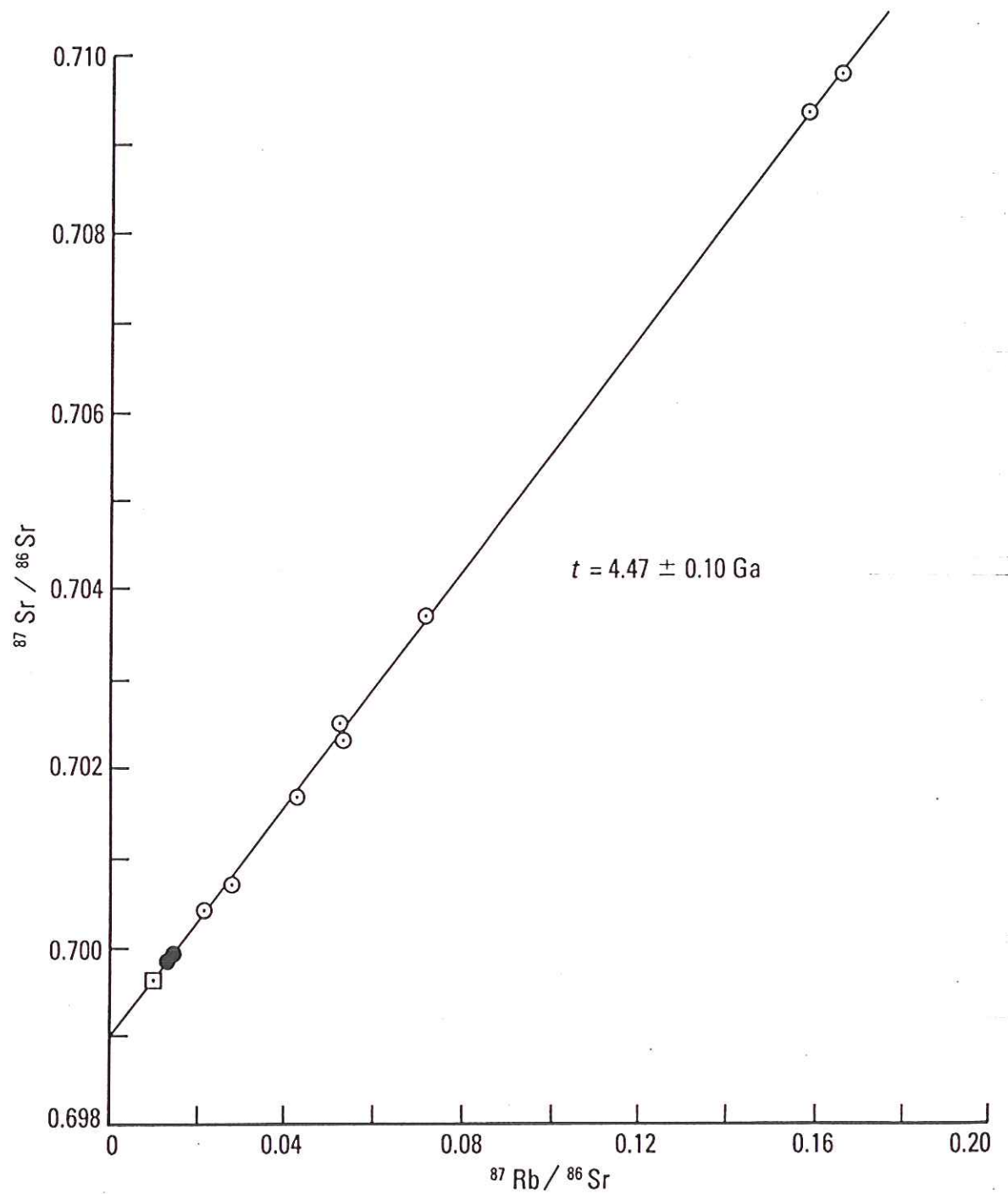


Fig. 5.18. Rb-Sr isochron for lunar dunite 72417. Open circles, chips of whole rock; filled circles, symplectites; square, olivine. All data were used in the isochron fit. (After Papanastassiou and Wasserburg, 1975.)

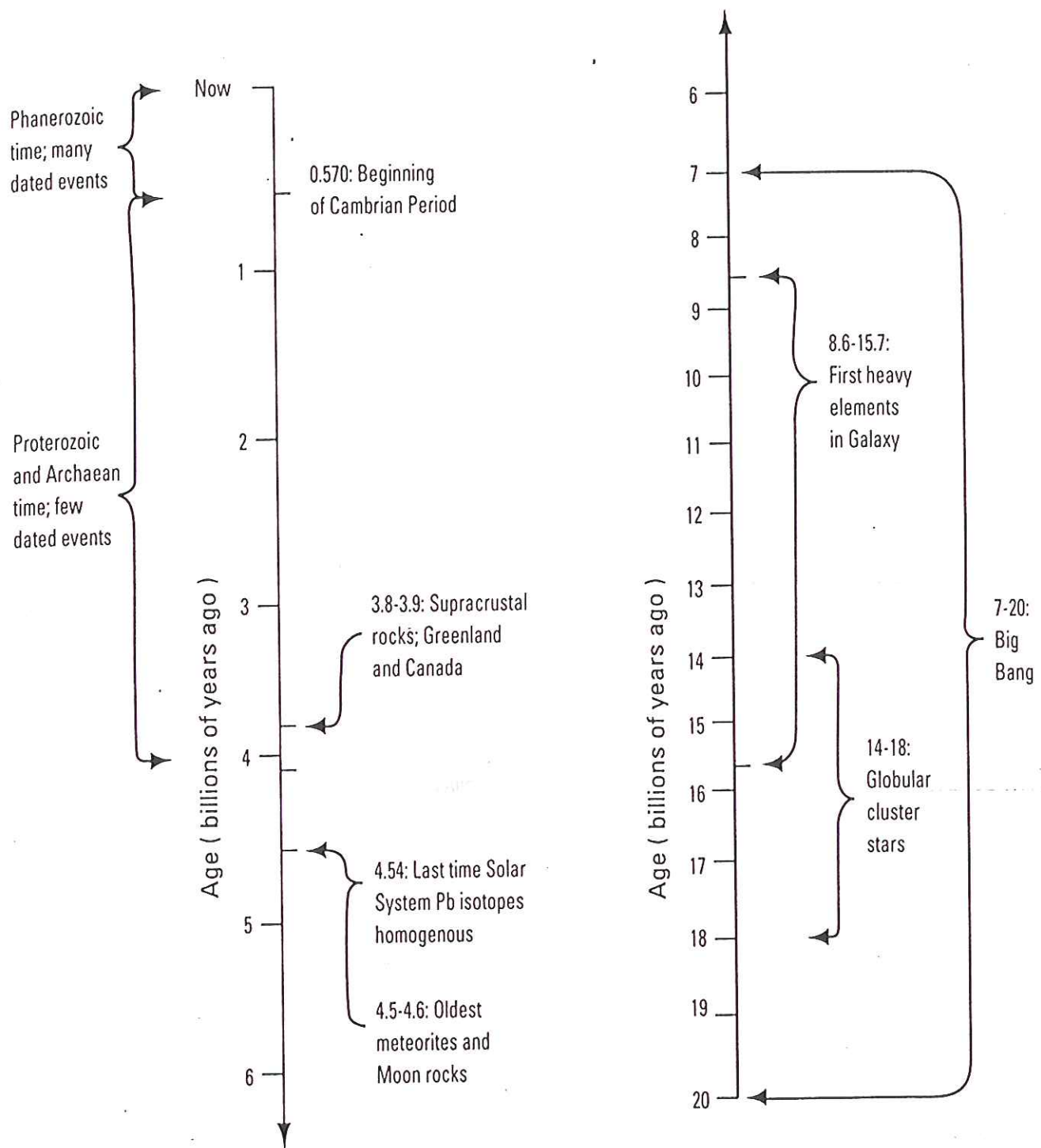


Fig. 1.2. Linear time scale of some important and datable events in the history of the Earth, the Solar System, the Galaxy, and the universe. The bases for the ages are discussed in chapters 4 through 8.

More modern value of Hubble constant
 → Big Bang was 12-13 b.y. ago

extra from
here on

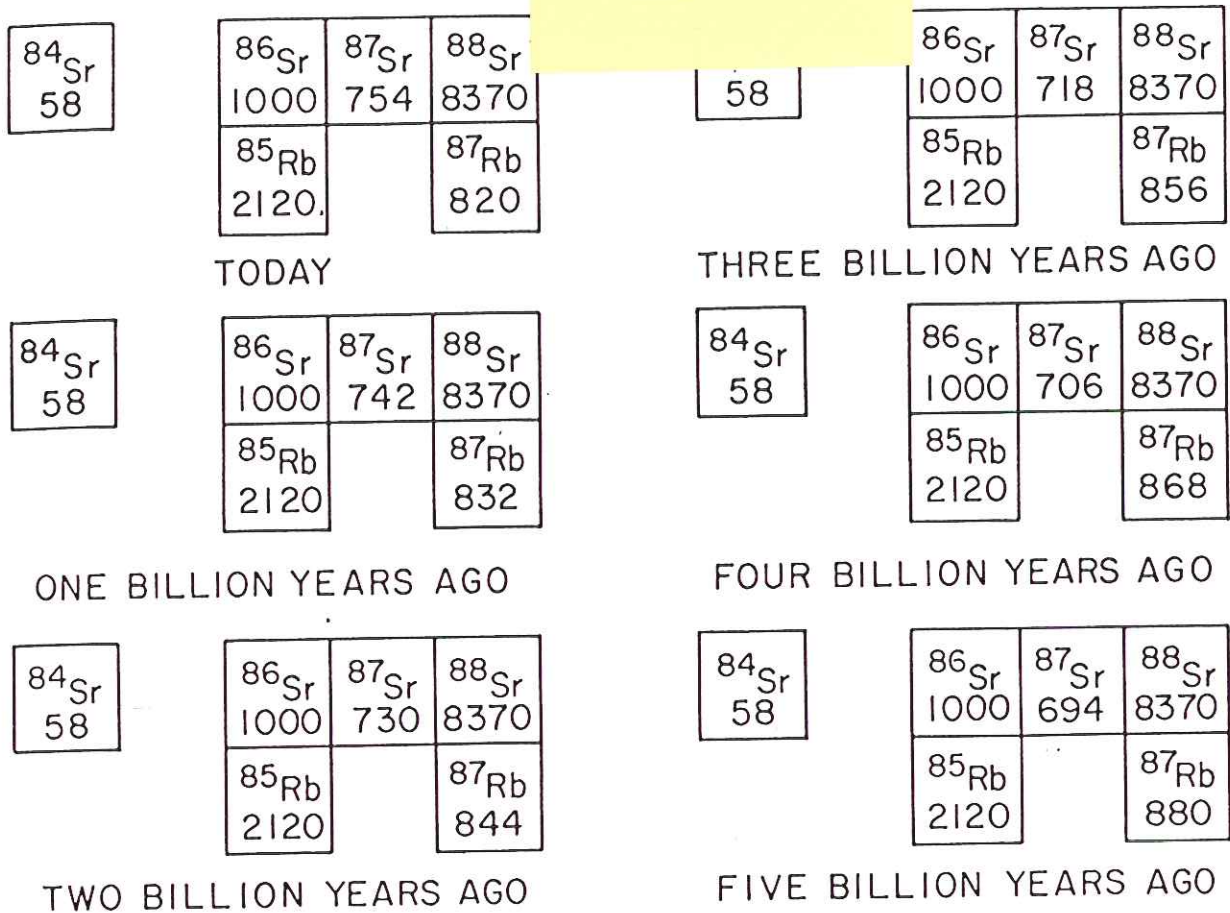
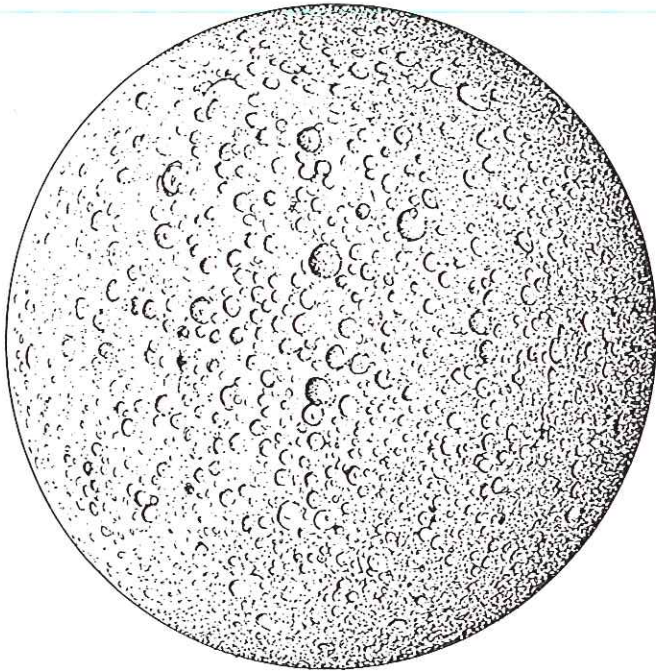


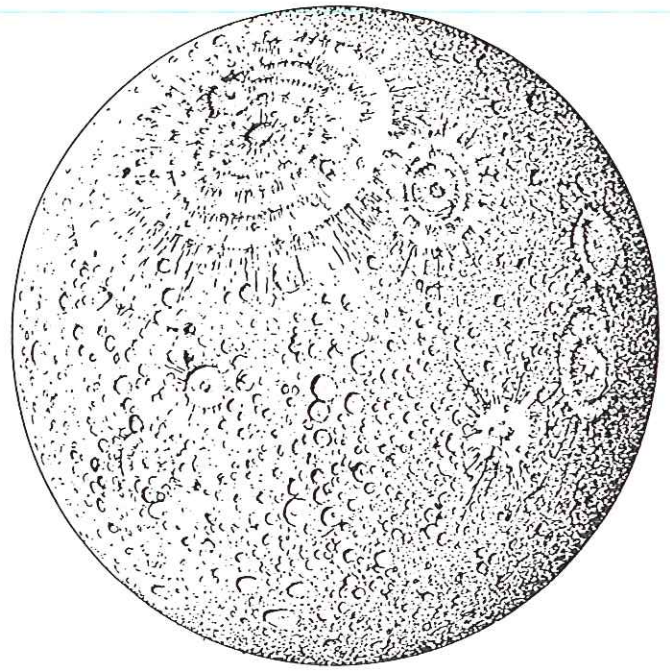
Figure 4-1. Evolution of the isotopic compositions of Sr and Rb in chondrites: The boxes in each of the six groupings on this diagram represent the naturally existing isotopes of elements strontium and rubidium. They are arranged as they would be on a chart of the nuclides. In the upper left grouping are shown the relative (to 1000 atoms of ^{86}Sr) amounts of these nuclides found today in a typical chondritic meteorite. Using the half-life of ^{87}Rb (49 billion years), it is possible to calculate the amount of ^{87}Rb in this chondritic meteorite 1, 2, 3, 4, and 5 billion years ago. Since one ^{87}Sr atom was produced for each ^{87}Rb atom that decayed, it is also possible to calculate how many ^{87}Sr atoms were present at each of these times. Since the remaining four isotopes are not radioactive and are not produced by the radiodecay of any other nuclide, their amounts do not change with time. The purpose of this exercise is to show that if the age of meteorites is in the range from 1 billion years to even 10 billion years, no more than about 15 percent of their ^{87}Sr atoms can have come from radiodecay. Most of the ^{87}Sr must have been there when the meteorite formed!



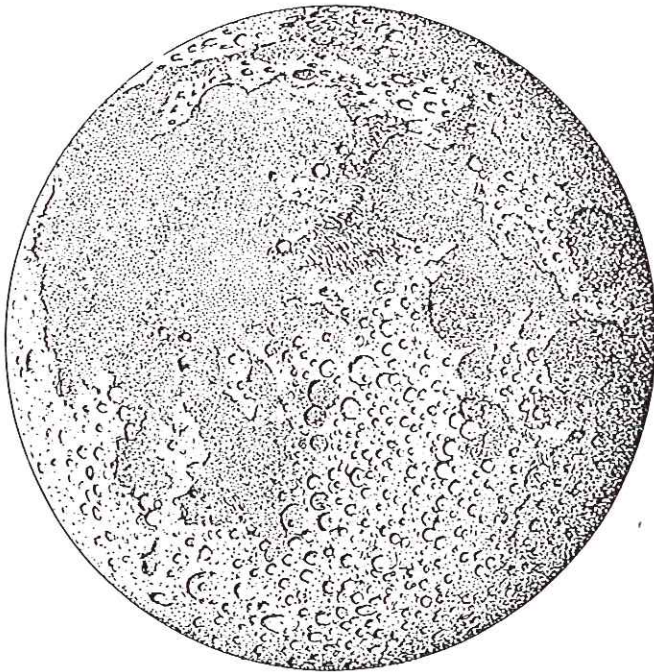
Fig. 5.1. The near side of the Moon, photograph (above) and diagram (opposite), showing some of the principal named features and the locations of the six Apollo (A) and three Luna (L) landing sites from which samples were returned to Earth. The dark smooth areas are maria; the brighter rugged areas are terrae (highlands). (Lick Observatory Photograph L-9, reproduced with permission.)



(A) Stage I. Formation of Moon by accretion (about 4.55 billion years ago) created densely cratered terrain over the entire surface of the Moon. The outer layers of the Moon may have been completely molten before this surface was shaped.



(B) Stage II. The formation of multiring basins (Imbrium Basin formed 3.9 billion years ago) is attributed to the impact of asteroid-sized bodies. The infall of these meteoritic bodies may represent the final stages of accretion. Remnants of this cratered surface are preserved in the lunar highlands.



(C) Stage III. Extrusion of the mare basalts (from about 4 billion to perhaps 2.5 billion years ago) was a manifestation of a major thermal event in lunar history, which occurred when the lithosphere was still relatively thin. Lava flows filled many of the multiring basins on the Moon's near side and in some areas they covered parts of the highlands that lack obvious multiring structures.



(D) Stage IV. Relatively light meteorite bombardment (from 3.2 billion years ago to the present) formed some craters with bright rays, but the rate of cratering has been greatly reduced. The lunar landscape has changed little during the last 3 billion years.

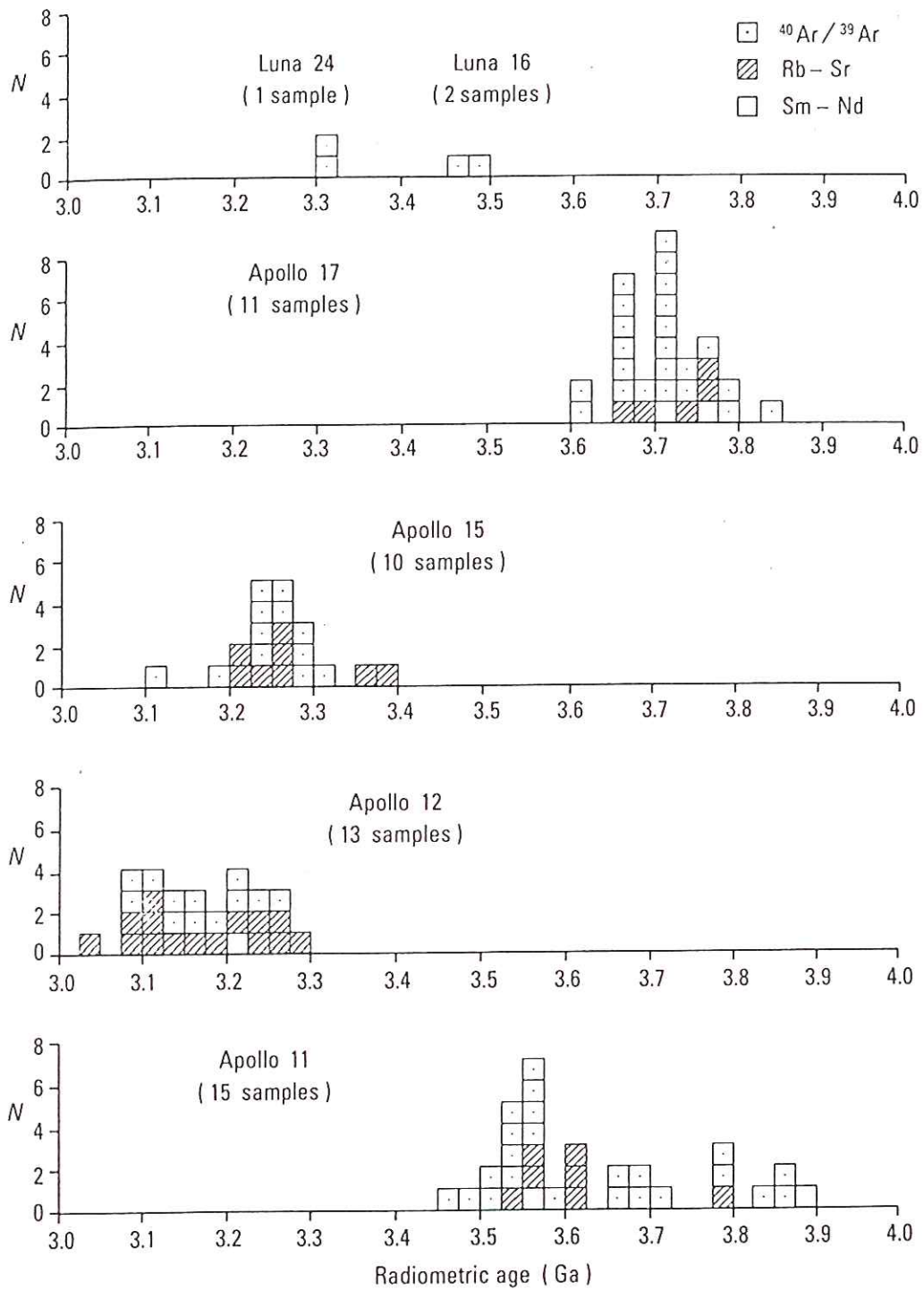


Fig. 5.10. Data on 114 radiometric age measurements on 52 samples of mare basalt. Included are ages whose reported analytical uncertainties at the 95% confidence level (two standard deviations) are 0.1 Ga or less.

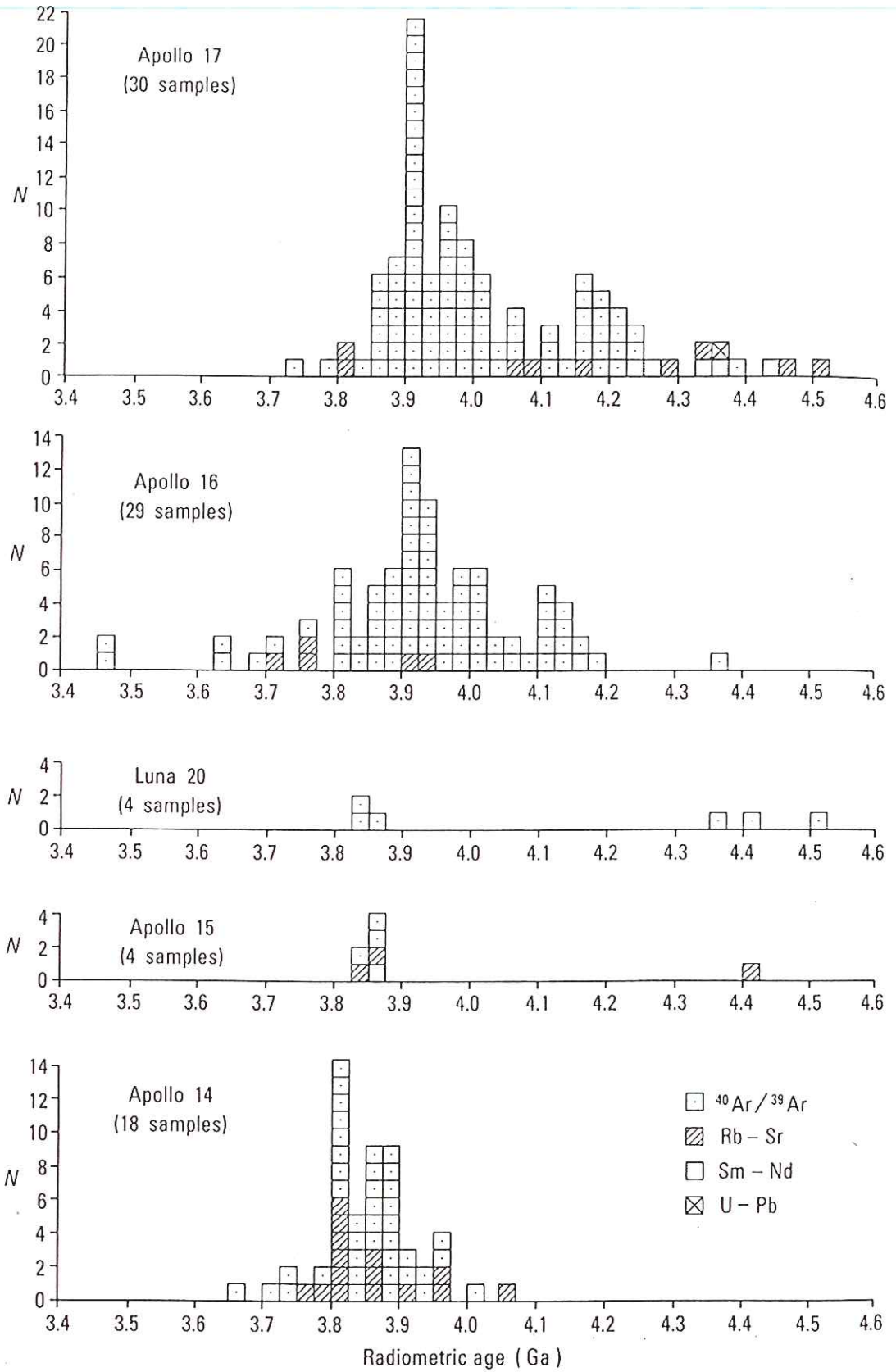


Fig. 5.12. Data on 260 radiometric age measurements on 85 samples from the lunar highlands. The ages included have analytical uncertainties of 0.1 Ga or less at the 95% level of confidence (two standard deviations).



FIG. 4-14 Tectonic provinces of the Canadian shield. (After C. H. Stockwell, Geol. Surv. Canada, Dept. Mines and Technical Surveys, Paper 63-17, 1963.)

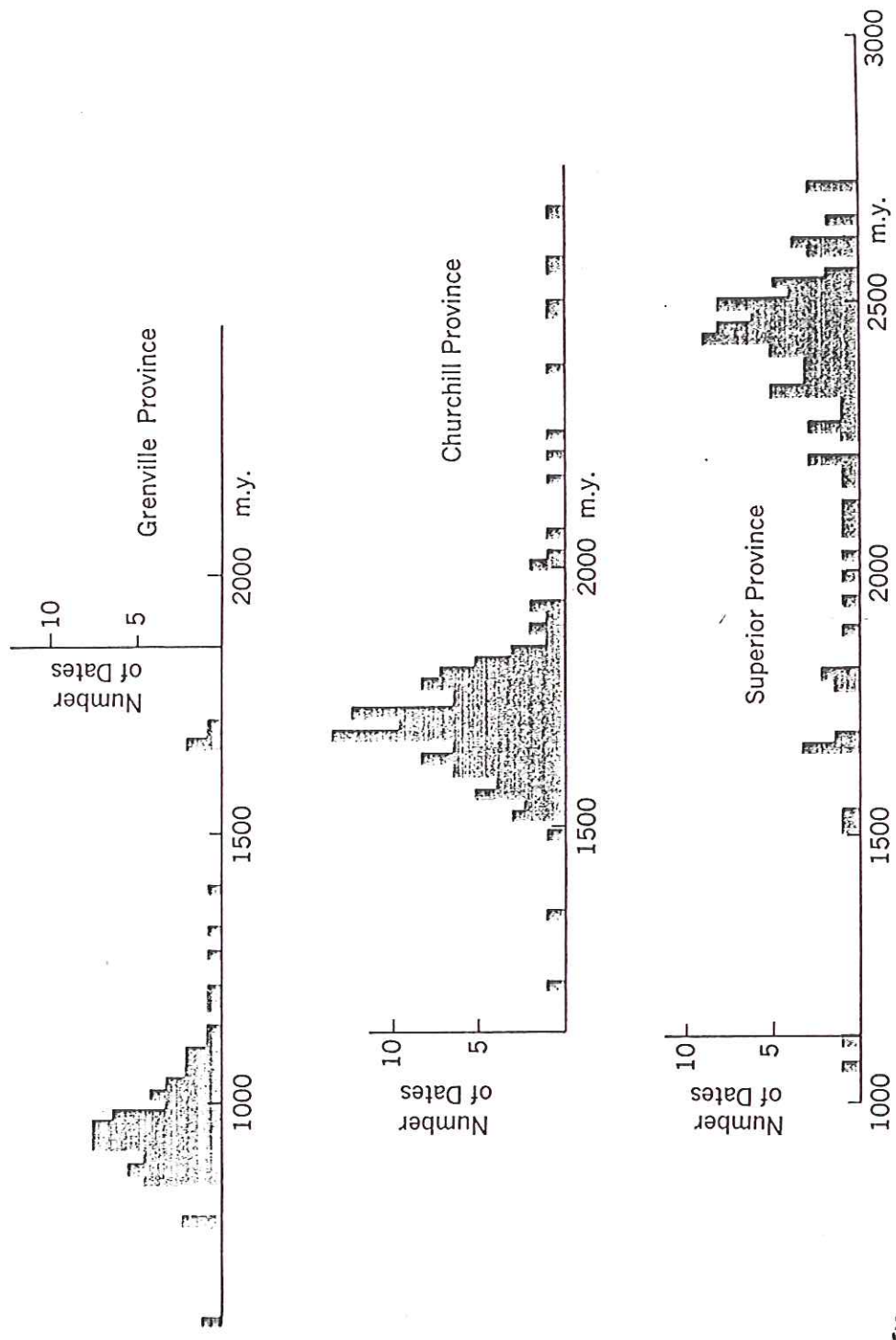


FIG. 4-15 Histograms showing numbers of dates determined by K-Ar analyses of micas from three tectonic provinces at the Canadian shield. (After J. A. Lowdon, Geol. Surv. Canada, Dept. Mines and Technical Surveys, Paper 63-17, 1963.)

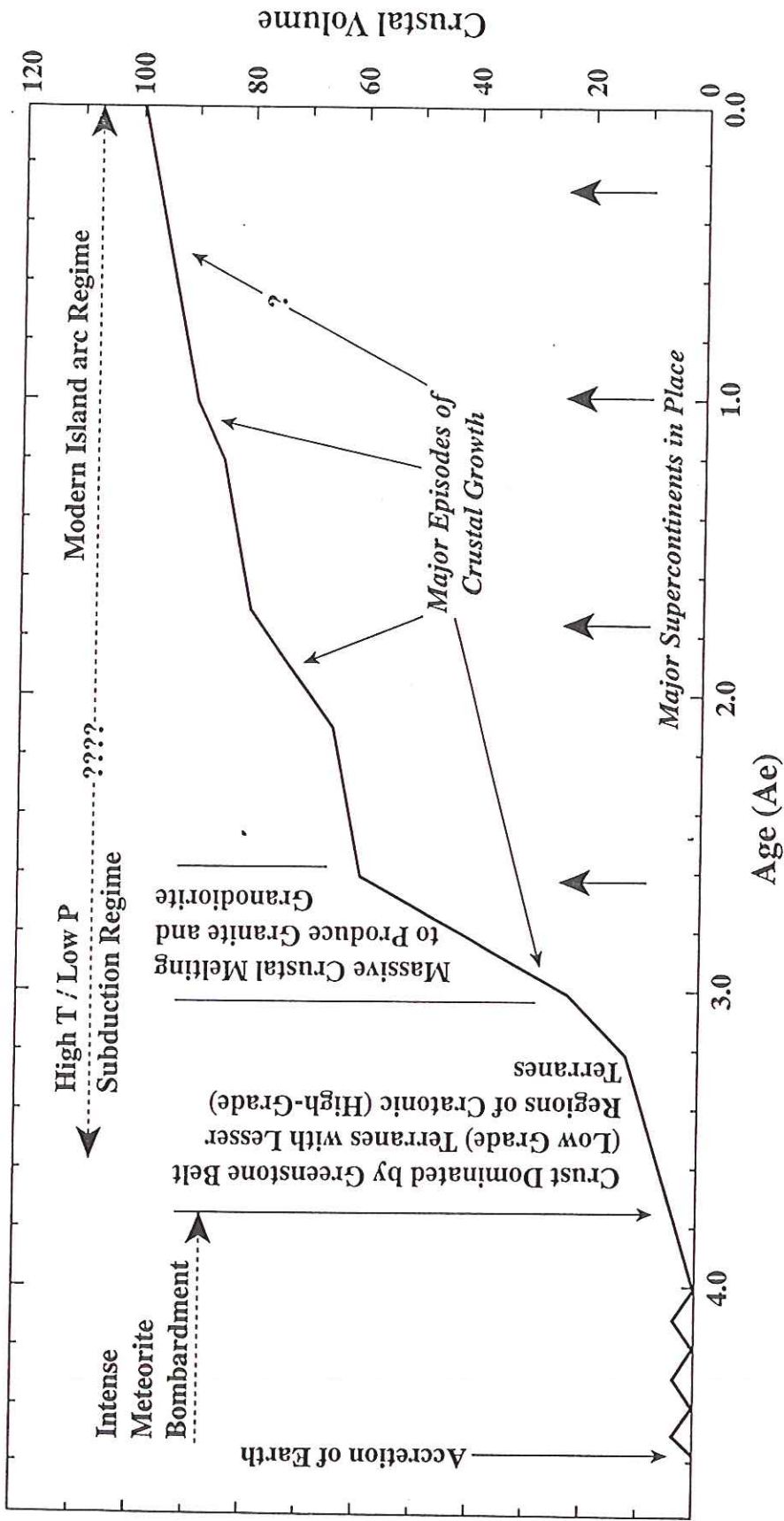


Figure 16.5. One view of the growth of continental volume, relative to its present value, over time. Some other major events in Earth's crustal history are shown, as well as a rough guess as to when modern arc volcanism came into play. From Taylor and McLennan (1995).

take K-spar
sample
one in every
5000 K atoms
is ^{87}Rb

single
2-sided
stray sheet