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I. INTRODUCTION

Meteorites can contribute in two ways toward a better understanding of the early history of the solar system. First, they seem to provide a more representative sample of average planetary matter than the highly-differentiated crust of the earth. Second, the meteorite parent bodies ceased to be geochemically active very shortly after their formation, some 4.5×10^9 years ago, and the meteorites thus provide a much better preserved record of the earliest events in the formation of the planetary system than does the earth.

Unfortunately, there is no consensus on the nature of the meteorite parent bodies, not even on such basic properties as size, location, and multiplicity (Table I). It is not too surprising that such disagreement exists: the meteorite samples commonly available for study represent only about 10^{-23} to 10^{-26} of the parent body, and some degree of personal bias is

* The material in this chapter is drawn in part from two papers by the author (Anders 1962, 1963).

inevitable in such a long extrapolation. Given this difference in models, even a simple observation, such as the fact that the iron meteorites were once molten, may occasionally lead to radically different interpretations. Ringwood (1959), Lovering (1957), and Wood (1958) assumed that the melting was caused by long-lived radioactivities (K^{40} , Th^{232} , U^{235} , and U^{238}); Urey attributed it to chemical reactions among thermodynamically unstable compounds (Urey and Donn, 1956) or adiabatic compression of gases (Urey 1962a); and Fish, Goles, and Anders (1960), following an earlier suggestion by Urey (1955), proposed that extinct radioactivities, specifically 740,000-year Al^{26} , were responsible for the melting. Not many properties of meteorites, however well established, permit only a single, unambiguous interpretation. Hence, the most one can demand of any theory of the origin of the solar system is that it be compatible with the primary properties of the meteorites, and with those secondary deductions on which a consensus exists.

It is not always possible to decide whether a given property reflects a process in the meteorite parent body or in the solar nebula. Even in those cases where the property clearly dates back to the nebular stage, the original record has often been altered by later events in the meteorite parent body, or elsewhere. These limitations of meteoritic evidence must be borne in mind at all times.

A very comprehensive review of the subject was given by Urey (1957a) in Vol. 2 of Physics and Chemistry of the

Earth. This paper is still of fundamental importance, although certain ideas contained therein have been modified by Urey in later papers (Urey 1956, 1957b, 1958, 1959, 1962a). The present chapter is of narrower scope, and deals mainly with evidence obtained since 1957. In some cases, the conclusions reached differ from those of Urey, partly because of the introduction of new evidence, and partly because of differences in point of view.

II. EARLY THERMAL HISTORY OF METEORITIC MATTER

Some important clues can be obtained from the carbonaceous chondrites. They are quite similar to "ordinary" chondrites in bulk composition (Wiik 1956), except for their higher content of water (10-20% vs <0.5%) and carbon (1-5% vs <0.1%). Part of the carbon occurs in the form of organic compounds; a fact that has lately enhanced the popular appeal of these meteorites.

A. Mineralogy

The mineralogy of the carbonaceous chondrites is particularly revealing (Table II). The minerals found can be divided into three classes: conventional, "high-temperature" minerals; "characteristic" minerals, peculiar to this class of meteorites; and trace minerals. Estimated relative abundances are given as negative logarithms to the base 2.

One can prove rather convincingly that at least some

of the characteristic minerals are alteration products of the high-temperature minerals, rather than vice versa. X-ray diffraction and optical studies of composite grains of olivine and Murray F mineral (a hydrated silicate), show that the olivine sometimes occurs in thin parallel plates of the same crystallographic orientation, though the individual plates are separated by a thin layer of exceedingly fine-grained, randomly oriented Murray F mineral. The common orientation of the olivine plates can be understood only if single-crystal olivine served as the starting material for at least that part of the Murray F mineral which is more intimately associated with it (DuFresne and Anders 1962a).

Many of the other characteristic minerals, too, appear to be hydrated silicates. This fact, and particularly the occurrence of hydrated MgSO_4 in distinct veins, suggests that liquid water must once have acted on these meteorites, converting high-temperature minerals to the characteristic minerals. A detailed analysis shows that the observed mineral assemblage is very nearly in a state of chemical equilibrium, corresponding to a pH of 8-10, a reduction potential of ≤ -0.2 volts, and a temperature not far from 300°K (DuFresne and Anders 1962a; Nagy, Meinschein and Hennessey 1962). We shall first attempt to establish the origin of the high-temperature minerals that served as the starting material for the carbonaceous chondrites, and then discuss possible settings for this aqueous stage.

B. Ancestral Material of Carbonaceous Chondrites

Several alternative origins have been proposed for the high-temperature minerals. Edwards and Urey (1955) and Urey (1961) have pointed out that the carbonaceous chondrites have a variable, and frequently lower, content of Na and K than the ordinary chondrites. In the most extreme case, Nogoya, this depletion amounts to a factor of ~ 4 , relative to the ordinary chondrites. Urey therefore suggested that the carbonaceous chondrites were derived from the ordinary chondrites [specifically, the high-iron group, $\text{Fe/Si} \approx 0.85$, Urey and Craig (1953)], by an alteration process that depleted the alkalis while introducing S, C, and a few other elements in free or combined form.

This picture has become less satisfactory now that the abundances of various trace elements in meteorites have been determined. Most elements occur in meteorites in approximately their "cosmic" abundances, as given by the semi-empirical abundance curves of Suess and Urey (1956) and Cameron (1959a). Other trace elements, including most chalcophile ones, do not conform to this pattern. As discussed more fully in Sec. IVB, they occur in approximately their predicted abundances in carbonaceous chondrites, but are depleted by factors of up to 1,000 in ordinary chondrites (Fig. 1). If the carbonaceous chondrites were derived from ordinary chondrites, as suggested by Urey, one would have to assume that the depleted elements were somehow added to the carbonaceous chondrites during the alteration process. In that case, it would be a remarkable

coincidence if 6 of the 7 elements happened to be restored to just their cosmic abundances. (The seventh, mercury, may be exceptional because of its high volatility, but it should be noted that the point in Fig. 1 is based on a single measurement.)

The olivine in carbonaceous chondrites has a highly variable iron content, ranging in a single meteorite from 3 to 54 mol% FeO in a typical instance (Ringwood 1961), whereas it is of nearly constant composition in ordinary chondrites (Mason 1960). This factor, too, makes it difficult to derive carbonaceous chondrites from ordinary chondrites by any simple process.

Another clue comes from the presence of primordial noble gases in carbonaceous chondrites (Fig. 2). All meteorites contain noble gases produced by cosmic rays or the decay of long-lived radioactivities, but the carbonaceous chondrites also contain primordial noble gases that can be distinguished from cosmogenic or radiogenic noble gases by their isotopic and elemental composition (Stauffer 1961; Anders 1962). With the exception of He^4 and Ar^{40} , most of which is radiogenic, the noble gases in an ordinary chondrite are produced chiefly by the action of cosmic rays on iron, silicon and other stable elements in the meteorite. For example, the three neon isotopes are made in nearly equal amounts in this process (Eberhardt and Eberhardt 1961), whereas in primordial neon (represented by neon in the earth's atmosphere) the ratio $\text{Ne}^{20}/\text{Ne}^{21}/\text{Ne}^{22}$ is 90.8/0.26/8.9. The elemental ratios differ too, as can be seen in Fig. 2. The bulk of the primordial noble gases once associated with the

matter of the terrestrial planets and the asteroids seems to have been lost at a very early stage in the history of the solar system (Russell and Menzel 1934, Brown 1949a, Suess 1949), and it is surprising that appreciable amounts are still present in some meteorites.

The presence of these gases in a meteorite is not necessarily an assurance of its primitive character; in a few instances substantial amounts have been found in ordinary chondrites and even in such highly differentiated meteorites as achondrites and irons. In some cases at least the trapping of these gases must have occurred in the meteorite parent body, perhaps by shock (Fredriksson 1962) or by some other mechanism involving an "internal atmosphere" (DuFresne and Anders 1962b). But the very low abundance of these gases in most ordinary chondrites makes it difficult to derive the carbonaceous from the ordinary chondrites without some ad hoc mechanism for the re-introduction of the noble gases.

The trace element abundances, the variations in the olivine composition, and the primordial gas content are most easily explained by assuming that both carbonaceous chondrites and the ordinary chondrites were derived from still more primitive ancestral matter. Perhaps the most embarrassing requirement for this material is that some of it at least must have passed through an earlier, high-temperature stage without losing its primordial gases completely.

It is possible to accomplish this in the meteorite

parent body, but some special assumptions are required (DuFresne and Anders 1962b). A more attractive possibility is offered by Wood's (1958, 1962a) hypothesis, according to which planetary matter, expelled from the sun at high initial temperatures, cooled by adiabatic expansion so that progressive condensation could take place (see also Suess 1948; Levin 1958). The least volatile constituents would condense to high-temperature minerals (olivine, pyroxene, and nickel-iron). Material passing through the liquid zone of the phase diagram might condense to droplets (chondrules), whereas the remainder would condense directly from gas to solid. The particle size and mineralogical purity would depend on the relative rates of nucleation and growth, as well as on the cooling rate. Other substances, e.g. H_2O , NH_3 , and carbon compounds, would condense on further temperature drop.

In the presence of cosmic proportions of H_2 and H_2O , any exposed metallic iron would be converted to magnetite below $600^\circ K$ (Latimer 1950; Urey 1952a) and some hydrated silicates might form from the finely divided dust. Each of these substances could trap some of the surrounding primordial gas. The eventual accretion of the (now cold) dust into solid bodies, and the separation of the solids from the noncondensable gas would proceed along the path outlined by Urey (1952a, 1954, 1956, 1957a, 1958), Fish, Goles and Anders (1960), Fowler, Greenstein, and Hoyle (1962), or Cameron (1962).

Another opportunity for a brief high-temperature era

might occur during contraction of the proto-sun. The luminosity may have been as much as 10^3 times its present value when the sun had contracted to 10^2 times its present radius (see Chapter 3 of this book). Regardless of mechanism, an early high-temperature stage may be regarded, if not an essential boundary condition, so at least a desideratum for any theory of the origin of the solar system, inasmuch as it provides a convenient explanation for some properties of the meteorites.

Meteorites representing this hypothetical, primitive material have not been identified with certainty, though several possible candidates have been proposed (Renazzo, Wood 1962b; Ornans, DuFresne and Anders 1962a). Renazzo, in particular, contains a mineral assemblage that is far removed from chemical equilibrium: chondrules of forsterite (Mg_2SiO_4 nearly free of iron), surrounded by metal rims, and embedded in a carbonaceous matrix containing Fe_3O_4 and a very fine-grained silicate. The chondrules and the metal must have formed under strongly reducing conditions at high temperatures, whereas the coexistence of free carbon with highly oxidized magnetite suggests low-temperature conditions, which would explain both the oxidized state of the iron (Latimer 1950), and its failure to react with carbon. The presence of glass in the chondrules suggests a fast cooling rate, on a time scale of days or less.

All these properties are difficult to explain in terms of evolution in a meteorite parent body, but agree rather well with those expected for a conglomerate of primitive material

formed according to Wood's hypothesis. The presence of iron-rich olivine in some carbonaceous chondrites is harder to understand, since only iron-free olivine should form in the presence of cosmic proportions of hydrogen. This seems to require some special assumptions, such as a partial gas-solid separation at an early stage, and more than one heating and cooling cycle. Although these assumptions are esthetically objectionable because of their complexity, there is no a priori reason why the temperature and composition of the nebula should always have changed monotonically.

The other chondrites may have evolved from this primitive material along several paths. The carbonaceous chondrites could have been produced by the (preterrestrial) action of liquid water. The enstatite chondrites could have been derived from either the primitive material itself, or from the carbonaceous chondrites by strong heating in a closed system, so that neither the trace elements other than mercury nor the noble gases would be lost. Finally, the ordinary chondrites must have been processed still more drastically. The primordial noble gases and the trace elements were lost somehow, the silicates were recrystallized until their iron content became uniform, and carbon and its compounds were either lost or consumed by reactions with reducible substances (Fe_3O_4 , Fe^{2+} , etc.).

If such a high-temperature stage ever took place, then cometary matter, too, must have passed through it. This raises some new possibilities in regard to the mineral composition of

comets. In particular, the presence in comet tails of metal (or magnetite?) spherules, inferred from polarization and scattered light measurements (Liller 1960), is somewhat easier to understand if at least part of the cometary material had a high-temperature history, even though its final accretion occurred at low temperatures. This view gains further support from the discovery in cosmic dust of metal flakes with amorphous organic attachments. The fall dates of these particles seem to be correlated with several meteor showers of cometary origin (Parkin, Hunter, and Brownlow, 1962). Perhaps Herbig's (1961) suggestion that the carbonaceous chondrites were derived from comets should be reexamined in the light of this possibility.

It must be emphasized that the early high-temperature stage does not eliminate the need for a heat source in the meteorite parent bodies. As pointed out in the Introduction, such a heat source is needed to produce the iron meteorites, and to account for the compact textures and pronounced metamorphism of many chondrites (Wood 1962b). If the meteorites had originated in bodies of lunar size, then either chemical reactions of thermodynamically unstable compounds, or adiabatic compression of gases can be invoked, although some special assumptions are needed in each case. Both processes require growth of lunar-sized objects prior to the dissipation of the solar nebula, a rather narrow range of heating and cooling rates, and a closely timed breakup. On the other hand, if the meteorites are derived from asteroids, none of the above mechanisms is

effective, and heating by 0.74×10^6 year Al^{26} seems to be the principal remaining possibility.

C. Aqueous Stage

If the meteorites came from a planet-sized body, then the requirement of an aqueous stage is easy to satisfy. Planets of terrestrial size are able to hold water vapor gravitationally, and can maintain bodies of liquid water, from ponds to oceans. If the meteorites come from the Moon, some ad hoc assumption is required. One such process was recently discussed by Urey (1962b).

For the asteroids, the prospects are bleak: of all the parent bodies discussed, they are least likely to retain liquid water at their surfaces, owing to their small size and consequent low escape velocities. But there is a way in which they could retain liquid water in their interiors. If the asteroids were ever heated by an internal heat source (e.g. extinct radioactivity), their interior temperatures would increase with depth. The surface temperature of the body, being controlled by the amount of solar radiation reaching it, might be around $100\text{-}200^\circ\text{K}$. Further inward, the temperature would rise until the melting point of ice was reached. Liquid water could exist in this zone, down to a depth where the boiling point at the prevailing pressure was reached (Anders 1963). At a central temperature of 1900°K , some 5% of the volume of the body will contain liquid water.

Of course, the water will not last forever. Above the

zone of liquid water, there will be a permafrost zone (Gold 1961). This permafrost zone can serve to retain an "internal atmosphere" within the meteorite parent body, and may have played a role in the retention of primordial noble gases (DuFresne and Anders 1962a,b). Ice from this zone will evaporate at a rate determined by its vapor pressure (Watson, Murray, and Brown 1961). The vapor pressure depends on the temperature, which, in turn, depends mainly on the distance from the sun. For an asteroid of 100 km radius, with an initial water content of 10%, these times range from 2×10^6 y to more than 10^{10} years, depending on the distance from the sun. These times are certainly long enough to allow the formation of the characteristic minerals.

III. PRIMORDIAL NOBLE GASES

Apart from their usefulness in the identification of primitive matter, primordial noble gases in meteorites also shed some light on the separation of solids and gases that must have occurred at some time during the formation of the solar system.

About twenty meteorites, including representatives of all major classes, are known that contain substantial amounts of primordial noble gases. Some of these meteorites (irons and achondrites) are several steps removed from primitive matter, and must have undergone extensive chemical processing in the meteorite parent bodies before the gases were incorporated

somehow in their present locations. This implies that the meteorite parent bodies were able to store appreciable amounts of noble gases over extended periods, perhaps in an "internal atmosphere" (DuFresne and Anders 1962b, Anders 1962). Other meteorites, such as the carbonaceous chondrites, may have retained their gases in the sites in which they were originally trapped from the nebula, although they, too, may have released and reabsorbed the gases during residence in the parent body. No convincing argument can be made for the remaining meteorites (enstatite chondrites and some brecciated ordinary chondrites), but it seems likely that the final trapping occurred in the parent bodies as well, possibly under shock conditions (Fredriksson 1962).

In the majority of cases, the original distribution seems to have been altered by diffusion losses at some later stage of the meteorites' history. These effects should be minimal for meteorites with high He^4 content, since He diffuses more readily than Ne and the heavier noble gases. Data for 4 meteorites with the highest He^4 content are shown in Fig. 3. For ease of comparison, the depletion factors (defined as the ratio of observed to "cosmic" abundance) are plotted against mass number. Data for the Earth and the enstatite chondrite Abee have also been included.

In the case of the earth, Suess (1949) drew attention to the fact that Kr and Xe had virtually the same depletion factor of $\sim 10^{-7}$, whereas Ar and particularly Ne were depleted by larger factors. The data could be represented by the

equation*:

$$- \log \frac{N_{\text{ter}}}{N_{\text{cos}}} = 10e^{-0.045 M/m_1} + 7.1 \quad (1)$$

which is identical in form to the equation for mass fractionation during escape of a gas from a gravitational field:

$$- \ln \frac{N_t}{N_o} = tAe^{-Mgr/RT} \quad (2)$$

(N_{ter} = terrestrial abundance; N_{cos} = "cosmic" abundance; N_o = initial abundance; N_t = abundance at time t ; M = molecular weight; m_1 = atomic weight unit; A = constant; g = gravitational acceleration in escape layer; r = distance of escape layer from center of gravity; R = gas constant; T = temperature).

For gases of high M , the fractionation factor in Eq. (1) approaches the asymptotic value of $10^{-7.1}$. Suess therefore suggested that the gas loss from the earth took place in two stages: first, a general depletion (without mass fractionation) to $10^{-7.1}$ the initial value, followed by a second stage involving mass fractionation. He pointed out, however, that the small value of the exponent $Mgr/RT = 0.045 M/m_1$ presented some difficulties. With the present value for the earth's gravitational acceleration, no escape of neon, let alone heavier gases, is possible even assuming the extreme conditions of $T = 3000^\circ\text{K}$ and $r = 15,000$ km. Consequently, Suess argued that the escape

* Better agreement with the Suess and Urey (1956) abundances is obtained by changing the constant in Eq. (1) from 7.1 to 6.9. This change has been made in Fig. 3.

occurred either during a planetesimal stage preceding the formation of the earth, or during an era when the earth's period of rotation was so short that centrifugal force largely offset the gravitational force in the upper atmosphere.

This mechanism may or may not be valid for the Earth, but it is evident from Fig. 3 that most of the meteoritic data do not fit an exponential fractionation function of the form of Eq. (1). Two basic patterns seem to be present. The two achondrites, Kapoeta and Pesyanoe, show virtually no fractionation from Ar to Xe, if allowance is made for the fact that the Kr and Xe points are only upper limits. The lighter gases, He and Ne, are fractionated slightly, but, as argued below, the depletion factors show a much smaller mass dependence than expected for gravitational fractionation. The three chondrites show a progressively smaller depletion of the heavier noble gases, in qualitative accord with Eq. (1). But the depletion factors for the lighter gases are grossly inconsistent with a fractionation mechanism dependent on mass alone. Any such function with an exponent large enough to reproduce the observed fractionation of neon relative to xenon must, by necessity, predict a very large depletion of helium. For example, Eq. (1) predicts a depletion of He relative to Ne by a factor of 2×10^4 . The observed depletion in chondrites amounts to less than a factor of two in 2 out of 3 cases, Murray and Pantar. The third chondrite, Abee, shows a much greater depletion, consistent with gravitational fractionation, but it is not certain that the He

was not lost at a late stage.

The isotopic ratios also do not agree with the expected values for gravitational fractionation (Table III). Geiss (1957) has used Eq. (1) to calculate the neon and argon isotopic composition in primordial matter from the terrestrial isotopic ratios. With the exception of Goalpara, the observed meteoritic $\text{Ar}^{36}/\text{Ar}^{38}$ ratios in Table III are constant to $\sim 5\%$, and are exceedingly close to the atmospheric value of 5.35, much below the calculated primordial value of 7.87. In the case of neon, the ratio is more variable, but the agreement with the calculated primordial value is worse, if anything.

Nevertheless, it is of interest that the meteorites with $\text{Ne}^{20}/\text{Ar}^{38}$ ratios closest to the "cosmic" value of 61 also happen to have the highest $\text{Ne}^{20}/\text{Ne}^{22}$ ratios (Fig. 4). Evidently, some mass fractionation did take place during the loss of neon. The problem is, to find a process that would give the observed isotopic fractionation of neon, without, at the same time, producing large elemental fractionations between Ne and He.

One such process is diffusion. For a gas trapped in a crystal grain of radius a , the rate of escape depends on the diffusion parameter D/a^2 , where D , the diffusion coefficient, is given by

$$D = D_0 e^{-Q/RT} \quad (3)$$

(Q = activation energy; D_0 = a constant). Both Q and D_0 vary from one mineral and one gas to another, the heavier gases and

more densely packed minerals generally having larger values of Q . It is seen from Eq. (3) that the fractionation between two gases will be strongly temperature-dependent, any differences in D being accentuated at low temperatures. The diffusion mechanism can thus be invoked to account for almost any observed elemental fractionation, large or small, particularly if the gases are unevenly distributed among several minerals differing in retentivity and grain size.

Diffusion will also result in isotopic fractionations, since both D_0 and Q are mass-dependent. Zähringer (1962) suggested, citing the work of Lazarus (1960), that D should be proportional to $M^{-1/2}$. Assuming an initially uniform distribution of gas in spherical grains of radius a , the fraction f lost during time t is found to be

$$f = 1 - \sum_{n=1}^{\infty} \frac{6}{n^2 \pi^2} \exp \left[- \frac{n^2 \pi^2 D(M) t}{a^2} \right] \quad (4)$$

which simplifies to

$$f = \frac{6}{\pi^2} \exp \left[- \frac{\pi^2 D(M) t}{a^2} \right] \quad (5)$$

if f is large (>0.85). Setting $D(M) \propto M^{-1/2}$, the following isotopic fractionation factors are predicted (Table IV). It is seen that the observed constancy of $\pm 5\%$ in the $\text{Ar}^{36}/\text{Ar}^{38}$ ratio implies a retention of at least 10% of the initial argon, whereas the variation in $\text{Ne}^{20}/\text{Ne}^{22}$ ratio ($\sim 30\%$) implies a loss of all but 0.1% of the initial neon content. As pointed out by Zähringer, such variations in the degree of retention would result if the

diffusion constants of Ne and Ar differed by as little as a factor of 4. The experimentally observed ratios $D_{\text{Ne}}/D_{\text{Ar}}$ range from 3 for a carbonaceous chondrite at 950°K (Stauffer 1961) to 10^4 for the achondrite Kapoeta at low temperatures (Zähringer 1962).

To account for the lack of elemental fractionation between Ne and He, some additional assumption is required. If the gases were initially distributed between two types of sites differing in retentivity, the lighter gases might be largely lost from one type of site, while being retained in their original proportions in the other. In fact, even a difference in grain size might lead to much the same result (Zähringer 1962). The grain radius enters the diffusion parameter D/a^2 to the inverse second power, so that a doubling of the grain size is equivalent to a reduction in the diffusion constant by a factor of 4, which, as mentioned above, would cause a one hundred-fold increase in the fraction of gas retained. Neither of these assumptions seems particularly contrived, and it thus appears that the diffusion mechanism is able to account for all the elemental and isotopic fractionations observed in the meteorites and the earth.

Suess (1962) has pointed out that the meteoritic data in Fig. 3 could still be explained in terms of his model (loss of all but $\sim 10^{-7}$ of the initial gas by hydrodynamic flow, followed by gravitational fractionation of the remainder), if a small amount of unfractionated gas were retained or added. By this view, most of the atmospheric and meteoritic neon is unfractionated, and is close to primordial neon in isotopic composition.

Then the "primordial" $\text{Ne}^{20}/\text{Ne}^{22}$ ratio calculated from Eq. (1) has no quantitative significance, and its disagreement with the observed values does not rule out gravitational fractionation. For argon, however, the unfractionated component should be negligible, yet an $\text{Ar}^{36}/\text{Ar}^{38}$ ratio virtually identical to the terrestrial value is found in meteorites differing greatly in absolute and relative noble gas contents, including the two achondrites. This fact is difficult to reconcile with gravitational fractionation, although the possibility cannot be ruled out entirely for the terrestrial gases, at least. However, the diffusion mechanism proposed for the meteorites is also capable of accounting for the terrestrial fractionations. Hence the constraint that solid objects of appreciable size ($r > 1000$ km) must have accreted before the dissipation of the gases, no longer seems binding.

Another possibility to be considered is that the Earth's noble gases may be of secondary origin. Urey (1952a,b) suggested that an appreciable addition to the Earth's noble gas inventory might have been made by cometary material, whereas Cameron (1962) proposed that much of the Earth's xenon was of solar origin. Neither of these suggestions is incompatible with the conclusion reached above: that noble gas fractionation by diffusion appears to be an adequate explanation for both the meteoritic and terrestrial data.

Unfortunately, the flexibility of the diffusion mechanism, which was helpful in accounting for a complicated set of data, also makes it difficult to deduce an unambiguous history

from these data. One observation of fundamental importance is the presence of noble gases in nearly cosmic proportions in the two achondrites. This seems to imply that the initial trapping of the gas by solid particles occurred at an early stage, prior to any elemental fractionation of the individual gases. The further accretion of the particles to larger bodies must have occurred under low-temperature conditions, to avoid preferential loss of the lighter gases. This places a limit the order of 1 km/sec on the impact velocity of the accreting particles, and seems to suggest that the sun did not attain its high-luminosity stage during the era of accretion. After accretion of the meteorite parent bodies, the gases were stored without loss in some reservoir, while the primitive matter melted and differentiated into achondrites.

It is not quite clear how the gases were reintroduced into the meteorites from this reservoir, but it seems likely that shock was involved in some manner. Kapoeta consists of two distinct fractions, which, though identical in chemical composition, differ in color and primordial gas content (Signer and Suess 1962). Only the dark fraction, which is also rich in glass, contains primordial gases. It appears that the differences between the two fractions were produced by shock, and this process may also have introduced the noble gases. For Pesyanoe, a variety of other mechanisms have been proposed (DuFresne and Anders 1962b), but it is not unlikely that shock was in fact responsible.

The evidence from chondrites is more ambiguous. It is certainly striking that the curves for all three chondrites in Fig. 3 agree so well in the region Ar-Xe, any differences being confined to the lighter gases. Yet the three meteorites differ drastically in their chemical and thermal histories, belonging, as they do, to the three principal subclasses of chondrites (see Sec. II). The ordinary chondrite Pantar, which is farthest removed from primordial matter, shows the same non-uniform distribution of noble gases between two fractions as does Kápoeta, and it is again probable that the gases were introduced in their present locations by shock. This makes it necessary to assume a reservoir (perhaps an "internal atmosphere") where the gases were stored before their final incorporation in Pantar. Primordial gases are also found in small amounts in nearly all ordinary chondrites (Geiss and Hess 1958; Zähringer 1962), and, in substantially greater amounts, in the carbonaceous and enstatite chondrites.

It is thus necessary to assume that the meteorites either retained some of their initial endowment of primordial gases, or that they all experienced some process that reintroduced the gases from a reservoir. Very little is known about the physical location of the gases in these types of meteorites, and it is hence not possible to decide between these alternatives.

It is also not clear whether the gases in the chondrites were fractionated in the parent body or in the nebula. The similarity of the Ar-Xe fractionation may be taken to indicate

either a common gas reservoir in the parent body, or a common fractionation history of the primitive matter before accretion. The fractionation of the He and Ne could have occurred at a later stage in either case.

If the initial trapping of the gases had occurred under anything like equilibrium conditions, then the primordial gas content of primitive dust (as approximated by the present gas content of meteorites) might provide a crude lower limit to the ambient gas pressure at time of trapping. Inasmuch as the ordinary chondrites are by far the most abundant class of meteorites, their average primordial gas content should give a conservative lower limit to the gas content of primordial dust (Table V). Somewhat higher values are obtained for the other classes, and for a weighted average of the three subclasses of chondrites. If it is now assumed that the gases were dissolved in the silicate phase of the dust under conditions where Henry's law was obeyed, the equilibrium partial pressures in the gas phase can be calculated. Assuming a gas phase of primordial composition, and a solubility of all noble gases equal to that of He at 1500°-2000°C (2.84×10^{-3} cc STP/g at 1 atm. pressure; Gerling and Levskii 1956), one obtains the following partial pressures of H₂ (the most abundant gas) in the hypothetical gas phase from which the meteorite gases were trapped (Columns 4 and 5, Table V).

These pressures are surprisingly high, particularly those calculated from the xenon content. Even the lowest of these are attained in current models of the proto-sun only at distances

well within the orbit of Mercury (Chapter 3 of this book). Urey's (1962a) proto-lunar objects do provide pressures of the required order, but it is not certain whether they would also furnish suitable conditions for the development of other properties of the meteorites. It seems that one or several further assumptions will have to be made to account for the primordial gases in chondrites.

1) Trapping in silicates at high temperatures was not the principal mechanism.

2) Adsorption by the low-temperature condensate (Wood 1962a) may have played an important role, and would explain the tendency of noble gases to occur in meteorites rich in carbon or carbon compounds. Some enrichment of the heavier gases would occur under these conditions.

3) Trapping by gas hydrates may have been important (Miller 1961). This mechanism, too, would favor the heavier gases.

4) After initial trapping in the primary condensates, the gases may have been selectively lost by reheating in the very early stages of the accretion process. The degree of loss would vary with the size of the accreting objects. Very complex fractionation patterns might result if the accretion of primitive matter to the meteorite parent bodies proceeded through planetesimals of intermediate size (Cameron 1962, Fowler et al 1962).

Thus it is difficult to establish the time and the circumstances of the initial gas-solid separation. The evidence

from achondrites suggests non-selective trapping, presumably at an early stage. The evidence from chondrites, on the other hand, suggests selective trapping under circumstances favoring the enrichment of the heavier gases, or selective losses leading to the same result. No simple model can satisfy both these conditions, and it is thus necessary to assume a more complicated series of events. It can only be said that the gases were still present in cosmic proportions at the time of separation of the first solid particles, and that they were dispersed by the time the earth had grown to a large enough size to hold the gases gravitationally. This conclusion is not new.

IV. CHEMICAL FRACTIONATIONS IN CHONDRITES

The chondrites show certain compositional variations among themselves. If they were all derived from a single pool of primordial matter of initially uniform composition, each compositional difference must indicate some chemical fractionation process. A priori, it is not possible to decide whether a given fractionation took place in the nebula or in the parent body. But in some cases, evidence exists that favors one or the other of these alternatives.

A. Metal-Silicate Fractionation

In absolute terms, the largest such fractionation is that between iron and silicon. Urey and Craig (1953) drew attention to the fact that the chondrites fell into two groups (Fig. 5), differing in total Fe content (22 and 28%) and hence Fe/Si ratio (0.60 and 0.85). Later, Urey (1959) pointed out that the density of the moon (3.38-3.41 at low temperature and pressure) was somewhat lower than that of the chondrites of the low-iron group (3.51). Either the moon had the same low iron content as the L-group chondrites, and a water content of several percent which would account for the lower density, or else its iron content was lower still. Previously, Urey (1952a, 1957a) had noted that the variable density of the terrestrial planets could be attributed to differences in the proportions of iron and silicon.

All these lines of evidence point to a fractionation between iron and silicon, presumably as metal and silicate, respectively. Urey has suggested two such processes:

(1) Heating of asteroidal-sized planetesimals, accreted from primordial dust in a reducing atmosphere. Under these conditions, silicates would be volatilized preferentially, whereas the denser metal would sink to the interior (Urey 1952a).

(2) Production of metal and silicates in "primary"

objects of lunar size, followed by breakup into minute fragments. The brittle silicates are expected to break into smaller and less dense fragments which would be lost preferentially with the gas during the dissipation of the nebula (Urey 1956).

If Wood's hypothesis is correct, so that metal and silicate grains were already present in the nebula immediately after its separation from the sun (Sec. IIB), then Urey's second mechanism can be simplified. These particles could be fractionated directly, without going through an interim period of residence within lunar-sized objects. This fractionation might be enhanced further by the ferromagnetic properties of the metal grains (Cameron 1959b). Moreover, magnetic forces may also have aided in the accretion of the micron-sized dust to larger aggregates. Most of the silicate particles in the Orgueil carbonaceous chondrite contain minute inclusions of magnetite, and similar inclusions have been seen in chondrules from Kaba (Sztrókay, Tolnay, and Földvári-Vogl 1961) and other chondrites (Fredriksson 1962). This possibility should be taken into account in any estimates of accretion rates.

It seems rather more difficult to achieve this fractionation in the meteorite parent body. In principle, metal can be removed by melting and gravitational settling, but the present textures of the chondrites seem to have been

acquired at relatively low temperatures so that additional, complicating assumptions must be made about their history. Also, the density variations among the planets must then be explained by some other mechanism (e.g. reduction of silicon to the metal, Ringwood 1959, see also Urey 1960).

B. Trace Element Abundances

As already mentioned in Sec. IIB, the chalcophile trace elements (Bi, Cd, Hg, I, In, Pb, Sb, Te, and Tl) seem to have been fractionated much more severely in chondrites than iron, depletion factors of up to 1000 having been observed (Fig. 1). It is not immediately obvious whether this fractionation occurred in the nebula or the parent body. The most promising approach seems to be to find some common property of these elements that distinguishes them from other unfractionated trace elements, and then to look for a mechanism that would produce a fractionation based on this property.

Cursory inspection might suggest volatility as a common characteristic, but closer examination reveals several inconsistencies: Bi and Tl are depleted more strongly than the relatively volatile Hg and I, for example. A better correlation is found with geochemical character. As already pointed out by Suess (1947) on the basis of the early data of Goldschmidt (1937), most of these "deficient" elements are chalcophile. Still another trend may be noted in Fig. 6:

for the strongly depleted elements on the left, the boiling points are higher than the decomposition temperatures of the sulfides, whereas the reverse seems to be true for the less strongly depleted elements on the right. The highly volatile elements Hg, Cd, and I seem to occupy an intermediate position, and some ad hoc assumption may be required in their case. The position of Pb is uncertain, as the estimates of the cosmic abundance diverge widely (0.47, Suess and Urey 1956; 22, Cameron 1959a). In this graph, the abundance of Pb in carbonaceous chondrites was assumed to be equal to the cosmic abundance.

On the basis of these data, Fish et al (1960) and Anders and Goles (1961) proposed a cyclic process that might account for the observed depletion. In an internally heated asteroid, the first major phase to melt will be the FeS-Fe eutectic (M.P. 1260°K). Owing to its high density, it will flow toward the center, extracting chalcophile elements from the chondritic mantle rock. In its flow toward the core, the melt will experience higher and higher temperatures, causing the trace element sulfides to decompose. The more volatile elements will return to the mantle, cancelling, in part, the original depletion, whereas the less volatile ones (Bi, Tl) will reach the core, and become concentrated in a sulfide-phosphide zone.

This model has its weaknesses (Anders and Goles 1961),

and it may well be that the fractionation, while being based on the properties displayed in Fig. 6, took place in an altogether different setting, perhaps even in the solar nebula. It is also possible that some other property was actually of importance in the fractionation. For example, it can be seen from Vogel's (1961) survey of the ternary phase diagrams of 28 metals with iron and sulfur, that all 3 "deficient" elements represented in his study (Bi, Pb, and Tl) have two characteristics in common. Their sulfides are reduced to the metal by free iron, and the metals are immiscible with metallic iron. These properties set them apart from all other elements included in this study (except Ag, for which the degree of depletion in chondrites is not known, however). It may well be that the fractionation was actually based on these properties. At present, however, this part of the record remains undeciphered.

V. EXTINCT RADIOACTIVITY AND GENERAL ISOTOPIC ANOMALIES

One of the most significant clues to the early history of the solar system is extinct radioactivity. Brown (1947) predicted that the decay products of such radionuclides (with half-lives 10^6 - 10^8 years) might be detectable in the meteorites if they had formed shortly after a nucleogenetic event. After several unsuccessful searches (Wasserburg and Hayden 1955; Reynolds and Lipson 1957; see Kohman 1956 for a review), Reynolds (1960a) at last detected a 50% enrichment

of Xe^{129} in the Richardton chondrite. This enrichment was probably due to the one-time presence in the meteorite of 16.4×10^6 year I^{129} .

Not all of the 15 or so radioactivities in the half-life range 10^6 to 10^8 years (Fig. 7) are likely to be recognized in meteorites by way of their decay products. To leave detectable traces, an extinct radioactivity must have a high initial abundance, and a large geochemical fractionation factor from its decay product. It is also essential that the decay product have at least one non-radiogenic isotope, so that an isotopic rather than merely a chemical enrichment results. If these criteria are applied to the radionuclides in Fig. 7, only six remain:

A. Chronology of the Early Solar System

The most important information that can be obtained from extinct radioactivity is a relative age scale for meteorites, and, hopefully, planets and satellites. The following assumptions are made in the derivation of such an age scale.

1. At some time in the past the extinct radionuclide X^A was synthesized either in the Galaxy or in the solar nebula. (This nuclide decays with a mean life τ to the stable nuclide Y^A .)
2. At some later time t_0 , production of X^A ceased abruptly, and unsupported decay began. The initial amounts at t_0 were X_0^A and Y_0^A .
3. After the formation of solid objects, chemical fractionations between elements X and Y set in. If the last such fractionation occurred at time t , and the object remained a closed system thereafter, the one-time abundance of X^A can be related to the present-day abundance of Y^A . Defining $\Delta t_{XY} = t - t_0$, and using the symbols r = radiogenic and p = present, we write:

$$X_t^A = X_0^A e^{-(\Delta t_{XY}/\tau)} = Y_{r,p}^A = Y_p^A - Y_0^A - Y_{r,t}^A \quad (6)$$

For convenience, the abundances can be normalized to

the stable, non-radiogenic isotopes X^B and Y^C

$$\frac{X_t^A}{Y^C} = \frac{X_o^A}{X^B} \cdot \frac{X^B}{Y^C} e^{-(\Delta t_{XY}/\tau)} = \frac{Y_{r,p}^A}{Y^C} = \frac{Y_p^A - Y_t^A}{Y^C} \quad (7)$$

where

$$Y_t^A = Y_o^A + Y_{r,t}^A$$

Of course, this involves the tacit assumption that the solar nebula was isotopically homogeneous, and that no appreciable isotopic fractionations took place (other than those caused by radioactive decay).

In principle, this equation can be used to calculate Δt_{XY} (= "decay interval", Goles and Anders 1961, or "formation interval", Cameron 1962), but there are two practical difficulties in the application of this method. The amount of X^A at t_o (i.e. X_o^A/X^B) and the isotopic composition of Y at time t (specifically, Y_t^A/Y^C) must be known. The latter can hardly ever be determined exactly, since it requires, for every object to be dated, another object that separated from the solar nebula at the identical time t , with so high a ratio of Y to X that subsequent decay of X^A did not change the isotopic composition of Y appreciably. Under favorable circumstances (e.g. when X_o^A/Y_o^A is small, and X_t^A/Y_t^A is large), even a crude approximation of Y_t^A/Y^C will suffice, but when the isotopic anomaly is small, the uncertainties become quite

large. To stress the distinction between Y_0^A/Y^C , which may be regarded constant for the entire solar system,* and Y_t^A/Y^C , which varies with t , we shall use the terms "primordial" and "primeval" for these two quantities.

The amount of the extinct radionuclide present at t depends on the model assumed for nucleosynthesis. It is generally agreed that the elements were continuously synthesized in the Galaxy (Burbidge, Burbidge, Fowler and Hoyle 1957). If the further assumption is made that the rate of synthesis has remained constant with time (Wasserburg, Fowler, and Hoyle 1960), the amount of a stable nuclide X^B synthesized during time T (defined as the duration of nucleosynthesis prior to the formation of the solar system) equals

$$X^B = K_P T$$

where K_P is the production rate of X^B .

For a radioactive nuclide X^A , the corresponding equation is

$$X^A = K_A \tau (1 - e^{-T/\tau})$$

In the special case of an extinct radionuclide with $\tau \ll T$, this equation simplifies to

$$X^A \approx K_A \tau \quad (\text{for } \tau \ll T)$$

Hence the ratio X_0^A/X^B in equation (7) equals:

*

Some doubts have now arisen concerning the validity of this assumption (Sec. VB).

$$\frac{X_o^A}{X^B} = \frac{K_A}{K_B} \cdot \frac{T}{T} \quad (8)$$

and

$$\frac{Y_p^A - Y_t^A}{Y^C} = \frac{K_A}{K_B} \cdot \frac{X^B}{Y^C} \cdot \frac{T}{T} \cdot e^{-\frac{\Delta t_{XY}}{\tau}} \quad (9)$$

Solving for Δt_{XY} , we obtain:

$$\Delta t_{XY} = \tau \ln \left\{ \frac{Y^C}{Y_p^A - Y_t^A} \cdot \frac{K_A}{K_B} \cdot \frac{X^B}{Y^C} \cdot \frac{T}{T} \right\} \quad (10)$$

This model may be an oversimplification inasmuch as the rate of nucleosynthesis can hardly have been constant over the lifetime of the Galaxy. It seems likely that the rate declined with time, as more and more matter became locked up in small and slowly-evolving stars. To allow for this effect, Kohman (1961) proposed inclusion of another negative exponential term in Eq. (9), whereas Cameron (1962) assumed a more complicated model of galactic history. However, for the construction of a relative age scale the model of Wasserburg et al is quite adequate, as it provides a self-consistent set of Δt 's for each extinct radionuclide. From the degree of agreement of these sets of Δt 's one can infer the shortcomings of the model, and the modifications required to bring the Δt 's into accord.

I. I^{129}

The β^- emitter iodine-129 is the extinct radionuclide par excellence. Its half-life, 16.4×10^6 years (Kohman 1961), lies in a convenient range; its abundance is easy to estimate, and its decay product, Xe^{129} , should be easy to detect even at low levels, since the meteorites (and the Earth) seem to have retained very little primordial xenon. Table VI lists Xe^{129} contents of meteorites in which both the xenon and iodine have been determined.

On the basis of their xenon content, the meteorites fall into three classes. The "ordinary" (bronzite and hypersthene) chondrites have low xenon contents and zero to moderate Xe^{129} enrichment. The enstatite chondrites have intermediate xenon contents and very large Xe^{129} anomalies. Finally, the carbonaceous chondrites have very large xenon contents with only a slight, if any, Xe^{129} enrichment. The amount of radiogenic Xe^{129} in each meteorite is difficult to determine unambiguously. The composition of primeval xenon is not known and atmospheric xenon is therefore used to correct the observed mass spectrum for non-radiogenic Xe^{129*} .

* Fish and Goles (1962) have pointed out that the "primeval" xenon is probably identical with the "ambient" xenon in the internal atmosphere of the meteorite parent bodies. They proposed an ingenious method for deducing the isotopic composition of this ambient xenon by diffusion studies. Thus far the necessary measurements have been carried out on four

This practice is questionable in itself, but it also introduces a further ambiguity. The mass spectra of meteoritic and terrestrial xenon are incongruent at all mass numbers, not just at mass 129. Hence the apparent amount of radiogenic Xe^{129} varies with the choice of stable isotope Y^C used in the normalization. These points are more fully discussed elsewhere (Anders 1962). In the present paper, Xe^{130} is used for normalization (Kuroda 1960).

Assuming* that the Xe^{129} anomaly arose from the in situ decay of I^{129} , one can calculate I^{129} - Xe^{129} decay intervals for those meteorites whose xenon and iodine contents are known (Table VI). The following parameters are used in this calculation:

1. $T = 1 \times 10^{10}$ years.
2. $\text{Xe}_t^{129}/\text{Xe}^{130} = 6.48$ (mean of the ratio in Murray and Earth's atmosphere). The appropriateness of this ratio is open to question, but the error introduced is likely to be small at least for meteorites with large anomalies.

meteorites, and the observed values for $\text{Xe}^{129}/\text{Xe}^{132}$ show a spread of 30-40%. Evidently, the substitution of atmospheric xenon for primeval xenon in eq. (7) is not always justified.

* This assumption is probably valid. For a review of the evidence, and complete references, see Jeffery and Reynolds 1961b, Anders (1962), and Zähringer (1962).

3. $K_{129}/K_{127} = 1.0$

4. $\tau_{129} = 23.6 \text{ m.y.}$

The physical event dated by the I-Xe method is the onset of xenon retention by the meteorite-to-be. Judging from the limited evidence from heating experiments (Zähringer and Gentner 1961; Jeffery and Reynolds 1961a, b), it seems likely that retention of Xe^{129} could begin only after the temperature fell to some 300°K or below. The cooling time depends not only on the size of the meteorite parent body, but also on the depth of burial within the body. It is interesting that the cooling times implied by the Δt 's for both the ordinary and the enstatite chondrites indicate a radius near 200 km for the meteorite parent bodies (Fig. 8). A radius of 200-300 km was previously inferred from two independent lines of evidence: the relatively small difference ($\leq 0.5 \times 10^9 \text{ y}$) between the "solidification" ages (Pb-Pb, Rb-Sr) and most "gas retention" ages (K-Ar, U-He), and the maximum environmental temperature of chondrites consistent with Ar^{40} retention (Goles, Fish, and Anders 1960).

The enstatite chondrites appear to be older than the other classes of meteorites. Indarch is definitely younger than the other two, implying either an earlier origin, a larger parent body, or a more deep-seated location in the parent body. For the ordinary chondrites, only two values are available, which agree rather closely. Certainly a

larger statistical sample will be required before a detailed interpretation of these data can be attempted.

It is also very significant that the Sardis iron meteorite has a much larger decay interval than the chondrites just mentioned. It seems that Sardis began to retain radiogenic Xe^{129} some 130 - 190 m.y. later than the chondrites. This difference in relative ages is exactly the opposite of that expected from Urey's theory (1957b, 1958), according to which the chondrites consist of the debris of achondrites and irons, evolved in earlier, "primary" objects of lunar size. On the other hand, it is consistent with the view that the irons come from the cores of internally heated asteroids, which, because of their central location, would cool more slowly than the chondritic mantle (Fish et al 1960). Of course, it is possible to devise alternative, ad hoc histories involving larger bodies, but the complexity of these hypotheses grows with the amount of evidence to be explained. It is obviously essential to obtain further data on the relative ages of irons and stones.

For the Murray carbonaceous chondrite, a large ambiguity exists. The $\text{Xe}^{129}/\text{Xe}^{136}$ ratio in Murray is 3.28, compared to 2.98 for the atmosphere. Taking atmospheric xenon to represent primeval xenon congenetic with Murray, one can thus compute a radiogenic Xe^{129} content of $109 \pm 11 \times 10^{-11}$ cc STP, and a decay interval of 106^{+9}_{-6} m.y. However,

if the data are normalized to Xe^{130} (Kuroda 1960), the $\text{Xe}^{129}/\text{Xe}^{130}$ ratios are 6.48 and 6.49, corresponding to a radiogenic Xe^{129} content of $0 \pm 11 \times 10^{-11}$ cc STP, and a lower limit of ≥ 144 m.y. for the decay interval.

It can be argued (Anders 1962) that the Xe^{130} normalization is preferable. Then the lack of radiogenic Xe^{129} in Murray, and its low K-Ar age (1.9 or ≤ 2.77 AE, Stauffer 1961, Reynolds 1960b) would imply a long cooling time. However, a prolonged high-temperature era is difficult to reconcile with the presence of primordial noble gases in this meteorite (Reynolds 1960b, Stauffer 1961). A more satisfactory assumption would be that the radiogenic Ar^{40} and Xe^{129} are located in sites of lower retentivity than the primordial Ar^{36} , Xe^{124} , etc.. Then a very mild heating would cause the loss of the former, while permitting retention of the latter. In the case of a few other meteorites at least, such differences in retentivity have indeed been demonstrated for Ar^{36} and Ar^{40} (Geiss and Hess 1958; Zähringer and Gentner 1961) as well as for Xe^{129} and Xe^{132} (Jeffery and Reynolds 1961a). Whichever explanation is finally accepted for Murray, will probably apply to the other carbonaceous chondrites as well. Their xenon contents and isotopic compositions are very similar (Reynolds 1960c, Krummenacher, Merrihue, Pepin, and Reynolds 1962), as are their iodine contents (Goles and Anders 1962) and K-Ar ages (Stauffer 1961).

The I-Xe method can also provide information on the relative ages and formation conditions of individual constituents of the meteorites. Merrihue (1962) found that the chondrules in the Bruderheim chondrite were much richer in Xe^{129} than the meteorite as a whole, whereas their content of primordial xenon was much lower. Although a quantitative interpretation of these data cannot be made without further information on the iodine content of the chondrules, it is already apparent that these observations are in better accord with Wood's hypothesis than with any other. In Wood's model, the chondrules are early objects formed at high temperatures, where little primordial gas would be retained, while the matrix consists of fine dust condensed at lower temperatures, and is therefore probably richer in trapped gases. The differences in Xe^{129} content might reflect differences in iodine content, or in the grain size and retentivity of the iodine-bearing minerals. (See, however, Sec. VB.)

2. Plutonium-244

After the mass spectrum of meteoritic xenon became known, Kuroda (1960) pointed out that the differences between meteoritic and terrestrial xenon seemed to indicate an excess of the heavy isotopes Xe^{131} - Xe^{136} in the Earth's atmosphere. These isotopes happen to be the only ones which would be produced in substantial yield by fission of a heavy nuclide (Khlopin and Gerling 1948). Indeed, when the mass spectra were normalized to the shielded

isotope Xe^{130} instead of the unshielded Xe^{136} , the Δ_i 's* showed a striking qualitative resemblance to a typical spontaneous fission yield curve (U^{238} or Cm^{242}), although the anomaly at mass 132 corresponded to the improbably high fission yield of $\sim 12\%$ (Kuroda 1960).

Kuroda showed that the spontaneous fission of U^{238} during the past 4.5×10^9 y could not account for more than 2% of the observed effect, and suggested that spontaneous fission of extinct Pu^{244} (76 m.y.) and Cm^{247} (≥ 40 m.y.), or neutron-induced fission of U^{235} (0.71×10^9 y) might have produced the excess Xe^{131-6} . Indeed, the levels of Pu^{244} expected on the basis of the Wasserburg-Fowler-Hoyle model are more than sufficient to account for all the excess Xe^{131-6} now in the Earth's atmosphere.

Goles and Anders (1961) therefore argued that either the earth's atmosphere must be younger than the chondrites (Pu-Xe decay interval of the atmosphere = 290 m.y. as opposed to I-Xe intervals of 60-120 m.y. for the chondrites) or that the solar system received a smaller initial endowment of Pu^{244} than expected from the model of Wasserburg et al (1960). The former view receives some support from the previously mentioned fact that the Earth's atmosphere has virtually the same $\text{Xe}^{129}/\text{Xe}^{130}$ ratio as the Murray meteorite, although Murray has a thousand-fold higher xenon content per gram than the Earth. The identity

* Defined as $(\text{Xe}^i/\text{Xe}^{130})_T/(\text{Xe}^i/\text{Xe}^{130})_M$, where T and M stand for "terrestrial" and "meteoritic".

in $\text{Xe}^{129}/\text{Xe}^{130}$ ratios could, of course, be explained in terms of equal ages and fortuitously identical I/Xe ratios for the two objects, but the explanation in terms of a later origin of the Earth's atmosphere seems less contrived. Still, Cameron (1962) and Kuroda (1961) have shown that the apparent difference in age can be somewhat reduced by a different choice of parameters, and a generous allowance for possible errors in these parameters. Another possibility, to be discussed in Sec. VB, is that most of the I^{129} in the early solar system was not inherited from galactic nucleosynthesis, but was produced by "last-minute synthesis" (Fowler et al 1962) in the solar nebula. If the level of synthesis had been lower at 1 a.u. than at 2-5 a.u., then the low Xe^{129} content of the earth would merely reflect a smaller initial endowment of I^{129} and not a difference in age.

Thus far, no evidence has been obtained for fission-produced Xe^{131-6} in meteorites. The levels to be expected are quite low (10^{-11} - 10^{-12} cc STP/g) and the detection may be feasible only in meteorites with low primeval xenon content. However, a search for these minute amounts may well be worth the effort, since the fortuitous circumstance that two extinct radionuclides decay to isotopes of the same element, xenon, makes possible the development of a new dating method based on both I^{129} and Pu^{244} (Goles and Anders 1961). Assuming that

$t_{\text{IXe}} = t_{\text{PuXe}}$, i.e., that the iodine- and the plutonium-bearing minerals became retentive for xenon at about the same time, one can arrange the decay equations for the two radionuclides in such a way as to eliminate or determine various unknown quantities, including the primordial $\text{I}^{129}/\text{I}^{127}$ and $\text{Pu}^{244}/\text{U}^{238}$

ratios, etc. Comparison of these ratios with those expected for various models of nucleosynthesis might provide some vital clues to the nature of the last nuclear events preceding the formation of the solar system. Also, once these ratios are determined, absolute ages can be calculated for all objects containing radiogenic xenon. In principle, it may thus become possible to date the origin of planets, planetary atmospheres, satellites, and meteorites to an accuracy of a few million years. Of course, this method again presupposes that the initial I^{129}/I^{127} and Pu^{244}/U^{238} ratios were uniform throughout the entire solar system.

3. Pb^{205} and Pd^{107}

Two other extinct radionuclides have been looked for, $\sim 24 \times 10^6$ year Pb^{205} and 6.8×10^6 year Pd^{107} . In contrast to I^{129} and Pu^{244} , their decay products (Tl^{205} and Ag^{107}) are relatively non-volatile. Hence the fractionation between parent and daughter must have ceased at an earlier stage, presumably at the time of the last solidification of the meteoritic minerals. The temperature must have been near $1500^\circ K$ at this stage, compared to $\sim 300^\circ K$ for the I-Xe and Pu-Xe fractionations. No enrichment of Tl^{205} was detected in 6 meteorites, and in at least one case (Canyon Diablo) the Pb/Tl fractionation was large enough to permit the calculation of a meaningful lower limit of ≥ 100 m.y. to the decay interval (Anders and Stevens 1960). In contrast, the silver from 3 iron meteorites, including Canyon Diablo, showed Ag^{107} enrichments of 2-4% (Table VII; Murthy 1960, 1962a).

Allowing for some uncertainty in the Pd/Ag ratio, and the Ag^{107} content of primeval silver, these enrichments correspond to decay intervals around $(0 \pm 20) \times 10^6$ years.

To account for the difference between this result and the I-Xe decay interval of the Sardis iron in Table VI, it is merely necessary to assume that the cooling time of the cores of the meteorite parent bodies from $\sim 1500^\circ\text{K}$, where Pd/Ag fractionations ceased, to $\sim 300^\circ\text{K}$, where Xe^{129} retention began, was the order of 2×10^8 years. Such times are consistent with other estimates.

On the other hand, the difference between the Pd^{107} - Ag^{107} and Pb^{205} - Tl^{205} decay intervals calls for further special assumptions, such as an error in the estimate of the abundance of Pb^{205} at t_0 . Another possibility is the preferential production of Pd^{107} at a relatively late stage in the formation of the solar system (Anders 1961, Fowler, Greenstein, and Hoyle 1962, Murthy and Urey 1962). It is difficult, however, to conceive of an event that would produce Pd^{107} at levels of several percent the abundance of neighboring nuclides, without, at the same time, also raising the abundances of longer-lived extinct nuclides. Finally, one must consider the possibility that the measurements themselves may be in error. Data by Chakraborty, Rushing, Stevens, and Anders (1962), though of lower accuracy, show no perceptible Ag^{107} enrichment in a number of iron meteorites, including two of those studied by Murthy (Canyon Diablo and Toluca),

and the nickel-rich ataxite Piñon whose high Pd/Ag ratio (400-800) would lead one to expect an Ag^{107} anomaly of $\geq 20\%$ if it were con-genetic with Canyon Diablo. The reason for the discrepancies between the two sets of data is not known, but it would seem that no firm conclusions can be drawn in regard to the presence of ex-tinct Pd^{107} in meteorites.

B. General Isotopic Anomalies

Another clue of great potential significance is provided by the "general" isotopic anomalies in meteorites. Reynolds (1960a) noted that in addition to the anomaly at mass number 129, meteoritic and terrestrial xenon differed at other mass numbers as well (Fig. 9).

Differences at heavier mass numbers can probably be ex-plained in terms of a fission contribution to terrestrial xenon, as noted above. For the lighter mass numbers, three different interpretations have been proposed.

Goles and Anders (1961) normalize the data to Xe^{130} , find an excess of the lighter isotopes in meteoritic xenon and attrib-ute this to charged-particle induced spallation reactions in the solar nebula. Such reactions would be expected to produce the light, neutron-deficient isotopes in relatively high yields, and since these isotopes are normally rare, even a small incremental amount can cause a disproportionately large change in their re-lative abundance. The isotopic differences between meteoritic and terrestrial xenon are then attributed to the lower density of matter in the asteroidal belt, which may lead to higher levels

of particle acceleration and spallation.

A detailed model for charged-particle acceleration in the solar nebula was proposed by Fowler et al (1962), in an attempt to explain the abundance of D, Li, Be, and B, and their isotopic composition. They suggest that intense irradiation with 500 Mev protons and thermal neutrons took place at a time when most planetary and meteoritic matter was contained in meter-sized planetesimals. The dimensions of the planetesimals must have been such that on the average only some 10% of the material was strongly irradiated*.

Cameron (1962) normalizes the data to Xe^{128} , and considers terrestrial xenon to be deficient in the light isotopes. He attributes this deficiency to the addition of solar xenon to the Earth's atmosphere, assuming that solar xenon was once irradiated by 100 ev neutrons during the deuterium-burning stage of the sun's evolution. This irradiation would slightly deplete the isotopes 124, 126, 131, 134, and 136, while enhancing

* It now appears doubtful whether thermal neutron reactions occurred at anything like the levels postulated by Fowler et al. Murthy (1962b) has shown that the nuclides Sm^{149} and Gd^{157} , with thermal neutron capture cross sections of 40,800 and 160,000 barns, occur in exactly the same isotopic abundance in meteoritic and terrestrial matter. To explain this in terms of the above model, it is necessary to assume that the mixing ratios of irradiated and unirradiated material were exactly identical at 1 a.u. and at 2-5 a.u.

isotopes 128 and 130.

Krummenacher et al (1962) normalize the data to Xe^{136} and conclude that the isotopic differences can be explained by:

a) addition of fission-product xenon to the meteorites; b) strong mass-fractionation of terrestrial xenon, leading to progressive depletion of the lighter isotopes.

All three hypotheses can be made to account in some ad hoc manner for the observation that meteoritic krypton shows much smaller anomalies, the order of 2% or less. On this basis alone, none can be disqualified, though the plausibility of these ad hoc assumptions varies. However, two additional pieces of evidence have recently become available, that have a bearing on this question.

Umemoto (1962) has observed an enrichment of the lighter isotopes in meteoritic barium that is qualitatively very similar to that in xenon (Fig. 10). Barium is a non-volatile element and hence neither the mechanism of Cameron nor that of Krummenacher et al can be invoked to explain the anomalies without further ad hoc assumptions, since both depend on the fractionation or transport of gases.

Moreover, the chondrules in Bruderheim show a very strong enrichment of Xe^{124} and Xe^{126} (Merrihue 1962). When the chondrules are heated to progressively higher temperatures, the $\text{Xe}^{124}/\text{Xe}^{132}$ and $\text{Xe}^{126}/\text{Xe}^{132}$ ratios rise, until maximum values of 3.4 and 4.8 times the atmospheric ratio are reached at 1100°C. Apparently not only Xe^{129} , but also $\text{Xe}^{124,126}$ are present in the

meteorite as separate components, wholly distinct from the primordial component represented by Xe^{132} and the other heavy xenon isotopes. In contrast to Xe^{129} , the lighter xenon isotopes do not have a long-lived radioactive progenitor, and it seems necessary to conclude that the excess $\text{Xe}^{124,126}$ was generated in situ by some nuclear process, most probably spallation by charged particles.

Recent production by cosmic rays is not a satisfactory explanation. As already pointed out by Reynolds (1960b) for several other meteorites, the anomalous $\text{Xe}^{124,126}$ shows no correlation with the cosmic-ray exposure age. In the present instance, the level of cosmic-ray irradiation inferred from the lighter spallation products (He^3 , Ne^{21} , and Ar^{38}) is too low to account for more than a fraction of the Xe^{124} , Xe^{126} , and since the cosmic-ray exposure ages calculated from all three nuclides are essentially concordant (Table VIII), preferential diffusion loss of the lighter cosmogenic gases is ruled out. Hence an earlier era of irradiation must be postulated. Some arbitrary assumption will be required to account for the absence of the lighter spallation products from that era.

It is possible to deduce a few boundary conditions for this process. Owing to the short range of charged particles, the irradiation must have occurred while the meteoritic matter was dispersed in space as aggregates of no more than meter dimensions, and possibly after some of the nebular gas had been dissipated. To permit retention of xenon, the chondrules must not have been heated strongly after the irradiation, except,

perhaps, in a closed system. This implies either that the irradiation occurred after the high-luminosity stage of the sun, or that the chondrules had already accreted to larger, well-insulated bodies at the onset of the high-luminosity stage. Yet it is necessary to assume some heating, sufficient to expel the lighter noble gases, but not sufficient to homogenize the Xe^{132} and $\text{Xe}^{124,126}$. In principle, this could have happened during metamorphism in the meteorite parent body (Wood 1962b).

Perhaps the most troublesome requirement is that the metal particles in chondrites must have differentiated into kamacite and taenite at some time between the initial high-temperature stage and the final cooling of the meteorite parent bodies. This differentiation requires slow cooling through the temperature range $600^{\circ}\text{--}400^{\circ}\text{C}$. There is good evidence that this differentiation can proceed in situ (Anders 1960) and it may therefore have occurred in the immediate parent bodies of the chondrites, rather than in an earlier generation of "primary" objects (Urey 1956, 1959; Urey and Mayeda 1959). However, it is difficult to imagine how two separate xenon components could persist at the temperatures in question without complete redistribution and homogenization. Regardless which model is used, the conclusion seems inescapable that at least one of the two xenon components was introduced at a late stage, after the final cooling of meteoritic matter. A priori, there is no obvious way of deciding which component came last, but since it was shown in Sec. III that the primordial gases were definitely introduced

at a late stage in at least a few meteorites, and possibly in all others, one can tentatively assume that the $\text{Xe}^{124,126}$ was produced in meteoritic matter at an early stage, and remained associated with it during residence in the parent body, whereas the primordial Xe^{132} , etc., being concentrated preferentially in another fraction of the primitive condensate, was not incorporated in the meteoritic matter until after the end of the last high-temperature stage. Of course, some reason must be found why the fraction carrying the primordial gases was not also rich in $\text{Xe}^{124,126}$. One possibility is that the spallation-produced $\text{Xe}^{124,126}$ were lost by recoil from the micron-sized dust carrying adsorbed primordial gases. In the absence of any further clues, room for speculation on this question is almost unlimited.

Of course, if the excess $\text{Xe}^{124,126}$ was produced by charged-particle reactions, then some I^{129} should have been made also. The excess Xe^{126} is a factor of ~ 500 less abundant than the excess Xe^{129} (8×10^{-13} g Xe^{129} /g), but even the latter amount is minute compared to the cumulative abundance of all elements heavier than xenon (~ 100 ppm), and could be accounted for by a moderate level of spallation. The lower relative abundance of $\text{Xe}^{124,126}$ can be attributed to a partial diffusion loss at a time when most of the Xe^{129} was still in the form of the relatively non-volatile I^{129} . It is interesting that the model of Fowler et al (1962) predicts a spallation yield of 8×10^{-11} g I^{129} /g for the strongly irradiated portion of planetary matter. This is a factor of 100 higher than the observed Xe^{129} content of the Bruderheim chondrules.

An interesting consequence of such a spallation era is that a substantial amount of 0.74×10^6 y Al^{26} should be produced also. This nuclide would provide a suitable heat source for melting and differentiation in some of the meteorite parent bodies (Urey 1955; Fish et al 1960).

If the I^{129} was indeed produced by charged-particle reactions during some late stage in the evolution of the solar nebula, then there is no assurance that the level of spallation, and hence the yield of I^{129} , was constant throughout the nebula. Particle acceleration would be more efficient at the low densities prevailing in the asteroidal belt than at the high densities near the Earth's orbit. Then the surprisingly small amount of radiogenic Xe^{129} in terrestrial xenon (Anders and Goles 1961, Butler, Jeffery, Reynolds, and Wasserburg 1962) does not necessarily imply a younger age of the Earth, but, more likely, a smaller production of I^{129} in the region from which the Earth accreted. As already mentioned, such regional variations in the $\text{I}^{129}/\text{I}^{127}$ ratio would seriously limit the applicability of the dating methods discussed in Sec. VA.

Were it only for the Ba, Kr, and Xe data, the present evidence would seem to suggest some type of nuclear process that affected meteoritic matter to a greater extent than terrestrial matter. Such a process should have produced isotopic anomalies in many other elements as well. Earlier workers failed to detect isotopic differences in many elements isolated from meteorites (Brown 1949b; Schmitt, Mosen, Suffredini, Lasch, Sharp, and Olehy 1960; Reed, Kigoshi, and Turkevich 1960), but since the

differences appear to be the order of 2% or less for a non-volatile element, such as Ba, it is possible that many anomalies were missed due to insufficient sensitivity of the analytical methods employed.

Some of the most recent data are, however, inconsistent with the spallation mechanism, or, for that matter, any of the previously discussed mechanisms. Umemoto (1962) did not detect any anomalies in meteoritic cerium (an element only two atomic numbers above barium) although its lightest isotopes are only about twice as abundant as those of barium. Even more puzzling are the data on molybdenum (Murthy 1962c). Some iron meteorites appeared to contain normal molybdenum whereas others showed mass-dependent anomalies of up to 7%. In contrast to Ba and Xe, the light isotopes were depleted relative to their terrestrial abundances. It seems that the only way to account for all the data is to assume that the solar nebula was isotopically inhomogeneous, owing to incomplete mixing in its parent dust cloud. Even so, the persistence of gross isotopic inhomogeneities in as highly differentiated materials as iron meteorites calls for some further ad hoc assumptions.

Clearly, the general isotopic anomalies are of great potential value in delineating the nuclear history of the solar system. In principle, isotopic data are less ambiguous than chemical data, and can thus lead to firmer conclusions. At the present time, the data do not fit into a neat pattern, and it may be best to await a confirmation of some of these measurements before attempting a detailed interpretation.

VI. CONCLUSIONS

Urey (1962a) has recently outlined a series of possible events in the history of the solar system that would account for the evidence, while remaining in accord with physical possibilities. The events suggested are:

"(1) Early in its history the sun acquired a nebula whose plane was that of the ecliptic. This nebula may have resulted from a separation of the disc from the sun during the latter's evolution from a dust cloud, and this separation may have been aided by magnetic fields. Or, the sun may have captured the nebula by collision with a small dust cloud early in its history. I regard this latter suggestion as improbable.

(2) Within this solar nebula lunar-sized objects of the composition of the non-volatile fraction of cosmic matter were formed. The objects became heated to high temperatures, possibly by chemical heating or heating in a convection polytrope, and developed the characteristic kamacite and taenite iron-nickel phases of the chondritic meteorites.

(3) At least in that region from which the meteorites originated, these objects were broken into small fragments by some marked crushing action. These processes occurred 4.5 AE ago.

(4) There is some evidence that the lunar objects formed throughout the region of the terrestrial planets, and that they broke up into small fragments. The silicate

fraction was selectively lost from the region of the terrestrial planets, resulting in an increased iron content of all these planets and a variable mean density due to the different degrees of silicate loss.

(5) The gases were lost in the region of the terrestrial planets while the solid materials were at low temperatures.

(6) The solid objects collected into the terrestrial planets probably so slowly that their general temperatures were low. It is likely that these planets have never been completely melted.

(7) Uranus and Neptune have also lost their volatiles, but these included only such substances as are gaseous at low temperatures. This circumstance indicates that temperatures during the loss of gases were similar to those in the solar system at the present time.

(8) Jupiter and Saturn retained their volatiles. This indicates that their accumulation process proceeded more rapidly so that high gravitational fields, capable of holding gases, developed. The absence of a planet in the asteroidal belt may be due to the reverse process, i.e., dissipation proceeded more rapidly than accumulation, and most of the material in this region was lost.

(9) The moon may be one of the primary lunar-sized objects. Its size, mass and composition as indicated by its density is in accord with this view. It may also be that the other lunar-sized satellites of the solar system

are primary objects which have escaped break-up and capture during the accumulation of the planets.

(10) It is likely that radioactive heating, heating by an initially high-temperature sun, synthesis of some elements by high-energy protons and other processes, played some or essential roles in the history as well."

Except for points (2), (3), and (4), this outline is very much in accord with the views of the writer, and the evidence presented in this chapter. An alternative outline, covering the first five points, follows.

(1) The material comprising the solar nebula cooled rapidly from high initial temperatures, leading to the formation of chondrules, metal grains, and other solid particles.

(2) The nebula was irradiated by charged particles, the level of irradiation varying with distance from the sun. The material from which the chondrites formed was not outgassed strongly after this stage; hence the sun must have gone through its high-luminosity era before stage (2) or toward the end of stage (3).

(3) The planets and meteorite parent bodies accreted at low temperatures. Some time during stages (1) - (3) a fractionation of metal and silicate particles took place, based on differences in density, magnetic susceptibility, or brittleness. Some time during stages (1) - (3), before the terrestrial planets had grown to a large enough size to hold gases gravitationally, the gases were lost.

(4) The meteorite parent bodies were heated to a varying degree by a transient heat source of unknown nature (possibly an extinct radioactivity, such as Al^{26}); some to melting temperatures (permitting the formation of achondrites and irons) and others to temperatures near 500°C where metamorphism could take place and the metal particles in chondrites could differentiate to kamacite and taenite. After final cooling, some redistribution of noble gases took place, possibly by shock.

The differences between these two outlines are entirely due to subjective differences in judgment. The issues on which Urey's and the writer's views differ most widely are: whether there was an early high-temperature stage; whether lunar-sized objects played a special role in the development of the meteorites and planets; whether the metal particles in chondrites could have acquired their structures in situ; and whether iron meteorites formed as individual entities or as fragments of larger objects. As these issues are resolved by further experimental and theoretical work, the two outlines are likely to approach each other more and more closely.

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Table I

Properties of Meteorite Parent Bodies

	Lovering (1957)	Ringwood (1961)	Urey (1959)	Fish, Goles, Anders (1960) Wood (1958, 1962a)
Size	planetary	lunar	lunar	asteroidal
Location	2 - 5 a.u.	2 - 5 a.u.	1 a.u.	2 - 5 a.u.
Number	one	several	one	several

Table II
Mineralogy of Carbonaceous Chondrites
(DuFresne and Anders 1962)

	Orgueil	Ivuna	Haripura	Cold Bok	Mighei	Murray	Ornans	Lancé	Mokoia	
Wiik's Class	I	I	II	II	II	II	III	III	III	
Subclass	A	A	B	C	C	C	D	D	E	
		<u>HIGH-TEMPERATURE MINERALS</u>								
Clinopyroxene						3				
Olivine		?	3	3	1	1	0-1	0-1	1	
Alpha Iron			9		10	10	7	5		
Gamma Iron								10		
Magnetic Troilite	5	5					5	5		
		<u>"CHARACTERISTIC" MINERALS</u>								
Orgueil IM	1	1								
Magnetite	1	1	3	3	11*		?	?	2	
Murray F				1	1	1				
Haripura M										
Mokoia HT and SW									5	
Epsomite	3	3	<u>6</u>	<u>6</u>	<u>6</u>	<u>6</u>	<u>>16</u>	<u>10</u>	<u>6</u>	
Sulfur	<u>6</u>	<u>6</u>	<u>6</u>	<u>9</u>	<u>9</u>	<u>9</u>	<u>>20</u>	<u>13</u>	<u>6</u>	
		<u>TRACE MINERALS</u>								
Dolomite	9	8								
Breunnerite	10									
Pentlandite					11					

(continued on next page)

Table II, Continued.

*Trace associated with metallic iron.

Estimated abundances are given as negative logarithms of 2. Thus Mighei is about half olivine and half "Murray F" mineral, with mere traces of iron, pentlandite, magnetite, epsomite, and sulfur. Italicized values are of lower accuracy.

Table III

Primordial Noble Gases in Meteorites

Meteorite	Class	Ne ²⁰ (10 ⁻⁸ cc STP/g)	Ne ²⁰ Ar ³⁶	Ne ²⁰ Ne ²²	Ar ³⁶ Ar ³⁸	Ref.
Novo Urei	Ureilite	<1.3	<0.0055		5.36	a
Goalpara	"	<1.0	<0.026		5.88	a
Lancé	Carb. Ch.	3.3	0.019		5.36	a
Felix	" "	5.6	0.031		5.40	a
"	" "	4			5.5	b
Mighei	" "	12	0.20	8.3	5.4	b
Ivuna	" "	20.9	0.22	9.8	5.37	a
Cold Bokkeveld	" "	10.0	0.23	9.0	5.5	b
Abee	Enst. Ch.	6.3	0.25	9.3	5.50	a,b
Murray	Carb. Ch.	53.8	0.39	10.6	5.37	a
"	" "	39.7	0.58	9.7	5.5	b
"	" "	63.2	0.83	9.2	5.47	c
Orgueil	" "	46.4	0.69	11	5.6	b
Mokoia	" "	309	9.1	13.2	5.39	a
Pantar (dark)	Veined Ch.	394 (178)	16	12.4 (11.6)	5.2	d,e
Pesyanoe	Aubrite	2150	11.9	12.2	5.46	a,f
"	"	2060	15	12.5	5.4	b
Kapoeta	Achondrite	2220	24	13.5	5.2	b
Atmosphere			0.52	10.2	5.35	
"Cosmos"			61.4	23.0	7.87	g

REFERENCES TO TABLE III

- a. Stauffer (1961)
- b. Zähringer (1962)
- c. Reynolds (1961b)
- d. König, Keil, Hintenberger, Wlotzka and Begemann (1961)
- e. Reynolds, Merrihue and Pepin (1962)
- f. Gerling and Levskii (1956)
- g. Suess and Urey (1956)

Table IV

Isotopic Fractionation by Diffusion

(Zähringer 1962)

Fraction of heavy isotope remaining (%)	Relative isotopic ratio		
	He^3/He^4	$\text{Ne}^{20}/\text{Ne}^{22}$	$\text{Ar}^{36}/\text{Ar}^{38}$
100	1	1	1
10	0.76	0.91	0.95
1	0.53	0.81	0.90
0.1	0.37	0.72	0.84
0.01	0.22	0.63	-

Table V

Average Primordial Gas Contents of
Meteorites

Meteorite class	Primordial gas content 10^{-8} cc STP/g		Calculated partial pressure of H_2 in gas phase atm.	
	Ar ³⁶	Xe ¹³²	from Ar ³⁶	from Xe ¹³²
Ordinary chondrites	2.9	0.034	7.4	1.2×10^4
Carbonaceous chondrites	85	0.81	220	2.9×10^5
Enstatite chondrites	26	0.067	68	2.4×10^4
Weighted average of all chondrites	6.9	0.063	18	2.2×10^4
Achondrites	23	?	59	?

Table VI
 $I^{129}\text{-Xe}^{129}$ Decay Intervals

Object	Class	$\frac{\text{Xe}^{129}}{\text{Xe}^{130}}$	Xenon Content		I^{127} ppb	Δt 10^6 y	Ref.
			10^{-11} cc STP/g Total	Xe_r^{129}			
Abee*	Enst.Ch.	15.2	730	430	145	63	b,d,i
Indarch	" "	19.1	650	250	270	91	b,g
Indarch	" "		470	210	270	95	b,i
St.. Marks*	" "	20.0	360	170	70	68	b,d,i
Bruderheim	Chondrite	7.37	104	3.6	16	123	a,b
Richardton	"	8.99	104	10	28	113	b,e
Murray	Carb.Ch.	6.48	4300	≤ 22	230	≥ 144	b,f
Sardis troilite	Iron	8.54	42	3.3	3600	254	b,h
Atmosphere		6.49	65	(<17)	(65)	≥ 170	c

* For these meteorites, Xe_r^{129} was obtained from Zähringer's $\text{Xe}^{129}/\text{Xe}^{132}$ and Xe^{132} data, subtracting the atmospheric $\text{Xe}^{129}/\text{Xe}^{132}$ ratio. The $\text{Xe}^{129}/\text{Xe}^{130}$ ratio measured by Jeffery et al is not quite consistent with these figures, owing to differences in normalization and in the contamination level.

References - Table VI

- a. Clarke and Thode (1961)
- b. Goles and Anders (1960, 1962)
- c. Goles and Anders (1961)
- d. Jeffery and Reynolds (1961a)
- e. Reynolds (1960a)
- f. Reynolds (1960b)
- g. Reynolds (1960c)
- h. Reynolds, Merrihue, and Pepin (1962)
- i. Zähringer (1962)

Table VII

Isotopic Composition of Silver in Iron Meteorites

(Murthy 1960, 1962a)

Sample	$\frac{\text{Ag}^{107}}{\text{Ag}^{109}}$	Standard Dev.
Toluca troilite	1.087	± 0.004
Sikhote Alin	1.091	± 0.005
Canyon Diablo	1.107	± 0.003
Reagent AgNO_3	1.062-	± 0.002
	1.066	

Table VIII
Age Determinations on Bruderheim Chondrite

Method	Age 10^9 y	Ref.	Interpretation
$I^{129}-Xe^{129}$	0.123*	a,b	Cooling to $<300^\circ K$
$K^{40}-Ar^{40}$	1.6	c	Storage at $300^\circ K > T > 200^\circ K$ followed by abrupt temperature drop. (Breakup to $100\text{ km} > r > 10\text{m}?$)
$(U, Th)-He^4$	1.5	c	
H^3-He^3	0.024	c	Breakup to meter-sized fragments resulting in strong irradiation by cosmic rays
H^3-He^3	0.035	d	
$Ar^{39}-Ar^{38}$	0.026	d	
$Cl^{36}-Ar^{36}$	0.033	e	
$Na^{22}-Ne^{22}$	0.027	e	

a. Clarke and Thode (1961)

b. Goles and Anders (1961)

c. Signer (1961)

d. Fireman and DeFelice (1961)

e. Honda, Umemoto, and
Arnold (1961)

* The I-Xe decay interval is reckoned forward in time, from the cessation of nucleosynthesis, whereas all other ages are measured backward in time, starting from the present.

FIGURE CAPTIONS

Fig. 1. Trace element abundances in carbonaceous chondrites and ordinary chondrites. Although the ordinary chondrites are strongly depleted in these trace elements, they occur in carbonaceous chondrites in nearly their "cosmic" abundances. This suggests that carbonaceous chondrites are more closely related to primordial matter than the ordinary chondrites. [Data were taken from the following sources: Bi, Hg, Pb, and Tl, Reed, Kigoshi, and Turkevich (1960), and Ehmann and Huizenga (1959); Cd, Schmitt (1962); I and Te, Goles and Anders (1962); In, Schindewolf and Wahlgren (1960); Sb, Anders (1960)].

Fig. 2. Noble gases in an ordinary and a carbonaceous chondrite. In Holbrook, these gases (except for radiogenic He^4) are produced by cosmic-ray induced spallation reactions on iron and other stable nuclides. The three neon isotopes occur in nearly equal abundance. In Murray, the isotopic abundances of Ne and Ar resemble those in the Earth's atmosphere, suggesting that these gases, too, are of primordial origin. A small amount of cosmogenic gas is present in Murray, as indicated by the increased abundances of He^3 and Ne^{21} relative to their atmospheric abundances.

Fig. 3. Depletion factors for noble gases in meteorites and the earth (for references, see Table III). Solid circles, achondrites; open circles, chondrites. Upper limits are denoted by

Fig. 3. cont. horizontal bars. The broken line indicates Suess' fractionation function (eq. 1).

Fig. 4. Isotopic composition of primordial neon as a function of $\text{Ne}^{20}/\text{Ar}^{36}$ ratio (Adapted from Zähringer 1962). The $\text{Ne}^{20}/\text{Ne}^{22}$ decreases with the $\text{Ne}^{20}/\text{Ar}^{36}$ ratio, indicating that some isotopic fractionation took place during the elemental fractionation of Ne and Ar. The solid curve shows the isotopic fractionation expected during diffusion, assuming no argon loss, and a $M^{-\frac{1}{2}}$ dependence of the diffusion coefficient (Zähringer 1962). Better agreement with the data would have been obtained if some Ar loss had been assumed for the meteorites on the left.

Fig. 5. Distribution of iron between oxidized (silicate) and reduced (metal + troilite) phases in chondrites (Urey and Craig 1953). Two groups appear to be present, differing in degree of oxidation, and total iron content. Two lines of constant iron content are shown.

Fig. 6. Boiling points of chalcophile elements and their compounds. The depletion of these elements in ordinary chondrites is not a simple function of volatility. The most strongly depleted elements Tl, Bi, and In have boiling points higher than the decomposition temperature of their sulfides. The reverse seems to be true for the less strongly depleted elements Se, Te, and Zn on the right. The elements of very high volatility (Cd, Hg, and I) show an intermediate degree of depletion.

Fig. 7. Long-lived natural radioactivities. (Adapted from Kohman 1956). Arrows indicate lower limits to the half-life.

Fig. 8. Cooling times of asteroids (after Fish and Goles 1962). The times required for central and volume-average temperatures to fall from 700°K to 200°K were computed from the relations given by Allan and Jacobs (1956). The following assumptions were made:

- 1.) The only heat source during cooling was K^{40} ,
 $1.45 \times 10^{-6} \text{ cal g}^{-1} \text{ y}^{-1}$;
- 2.) thermal diffusivity $K \approx 1 \times 10^{-2} \text{ cm}^2 \text{ sec}^{-1}$;
- 3.) the surface temperature was fixed at 100°K by radiation.

Fig. 9. Isotopic composition of Murray xenon relative to atmospheric xenon. (After Reynolds 1960b). Lighter isotopes are present in excess.

Fig. 10. Isotopic composition of meteoritic barium relative to terrestrial barium (Umemoto 1962). The three meteorites are the ordinary chondrite Bruderheim and the calcium-rich achondrites Nuevo Laredo and Pasamonte.

Fig. 1

Fig.2

Fig.3

Fig.4

Fig.5

Fig.6

Fig.7

Fig. 8

Fig.9

Fig.10