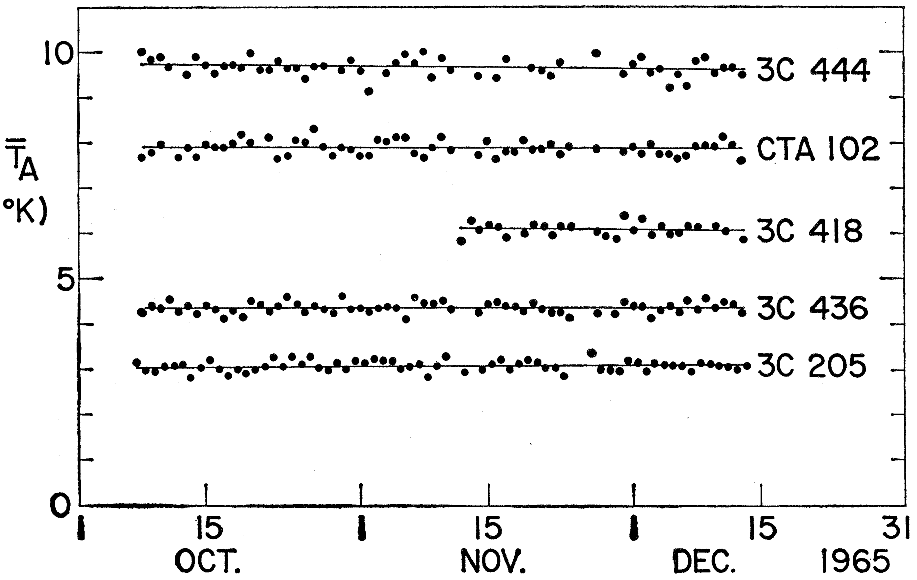

Fig. 1. Averages of antenna temperatures measured on two orthogonal polarizations plotted against date (U.T.) for the sources CTA 102, 3C 205, 3C 418, 3C 436, and 3C 444.

mean, and the least-square results indicate no linear change greater than 1.4 percent per 30 days. The standard deviation of the measurements about the least-square fit is 2 percent of the mean antenna temperature. We consider this proof that there were no significant changes in the intensity of CTA 102 during the 67-day observing period at 25.5-cm wavelength. Our observing period was not long enough to rule out completely changes of intensity of the type reported by Sholomitskii. However, Sholomitskii has shown that his intensity variations can be fitted by a sinusoidal law with a period of about 100 days and an amplitude of ± 23 percent, and our measurements would be sensitive to such variations. Our observations put an upper limit of about ± 4 percent on the amplitude of variations having a 100-day period. Based on Kellermann's flux-density scale (*8*) the observations correspond to a flux density for CTA 102 of $(7.2 \pm 0.4) \times 10^{-26}$ watt m^{-2} $(cy/sec)^{-1}$, where the error is estimated standard error.

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6 April 1966

Strontium Isotopes in Deep-Sea Sediments

Abstract. *Recent deep-sea sediments, leached of carbonate, have $Sr^{87}:Sr^{86}$ ratios ranging from 0.7044 to 0.7394. Strontium in the detrital sediment has not equilibrated isotopically with seawater strontium. Amounts of strontium-bearing authigenic material in the clay-mineral assemblage are not great enough to mask strontium-isotope ratios inherited from source areas.*

Isotopic studies have provided significant information about the geologic provenance of deep-sea sediments. Chow and Patterson (*1*), measuring $Pb^{206}:Pb^{207}$ ratios in lead, leached with HCl, from deep-sea sediments, found regional variations within the major ocean basins. They suggest that the lead is extracted biogenically from overlying water before complete oceanic mixing occurs; hence, soluble lead supplied to the oceans from each major drainage basin retains its isotopic identity in the marine realm.

Hurley *et al.* (*2*) obtained K-Ar "dates" from deep-sea sediments of the North Atlantic that range from 200 to 400 million years—ages compatible with the old ages of rocks in adjacent continental areas. A Pacific deep-sea sample "dated" at 80 million years reflects the generally younger rocks rimming the Pacific basin.

The purpose of investigating strontium-isotope ratios in deep-sea sediments is threefold: (i) to establish geologic provenance for deep-sea sediments, (ii) to distinguish between authigenic and detrital components of deep-sea sediments, and (iii) to explore the possibility of dating material recovered from deep penetrations of the marine sediment column, such as the Mohole, by the Rb-Sr method.

The $Sr^{87}:Sr^{86}$ ratios herein reported were obtained from the tops of cores raised by Lamont Geological Observatory. Chow and Patterson's lead-isotope provenance maps and Biscaye's (*3*) clay-mineral distribution data for Recent sediments were used to select the samples for this reconnaissance study. Biscaye's work indicates that continental weathering, rather than authigenic mineral formation, is the dominant factor controlling the mineralogy of Atlantic deep-sea sediments. Strontium-isotope differences in the detrital, aluminosilicate fraction of deep-sea sediments, therefore, must result from differences in provenance unless equilibration with sea-water strontium has occurred. The long residence time of strontium in the sea results in a nearly homogenous $Sr^{87}:Sr^{86}$ ratio of about 0.709 for sea water (*4*). If any component of the aluminosilicate fraction is authigenic, diagenetically altered, or otherwise modified isotopically by sea-water strontium, its presence will subdue differences due to provenance.

Calcium carbonate was dissolved from the sediments with a buffered solution of acetate and acetic acid (*5*). Calcium carbonate–rich sediments of the deep sea contain about 2200 parts per million strontium, whereas the noncarbonate or "clay" fraction contains about 50 to 250 ppm (*6*). Biogenically deposited calcium carbonate incorporates strontium that is isotopically identical with strontium in sea water; thus a small amount of calcium carbonate analyzed with the sediment

Table 1. Strontium-isotope composition of the Eimer and Amend standard $SrCO_3$, lot No. 492327. D, double rhenium filaments; S, single rhenium filament.

Date (Jan. 1966)	Ion source	$Sr^{87} : Sr^{86}$, (corr.)
19	D	0.7080
22	S	.7064
22	D	.7069
26	D	.7080
28	S	.7082

can mask the ratio in the detrital material.

Strontium was separated from the remaining silicate detritus by standard dissolution and ion-exchange procedures, and the ratios were determined on a 6-inch (15-cm) Nier-type mass spectrometer with both single and double rhenium-filament ion sources. The $Sr^{87} : Sr^{86}$ ratios were corrected for mass fractionation in the spectrometer by normalizing to a $Sr^{86} : Sr^{88}$ value of 0.1194. A strontium-isotope standard (*7*), analyzed five times during the study (Table 1), gave an average $Sr^{87} : Sr^{86}$ value of 0.7075, with a standard deviation of ±0.0007. The strontium-isotope ratios and other data obtained from the deep-sea cores appear in Table 2; the isotope ratios, in Fig. 1.

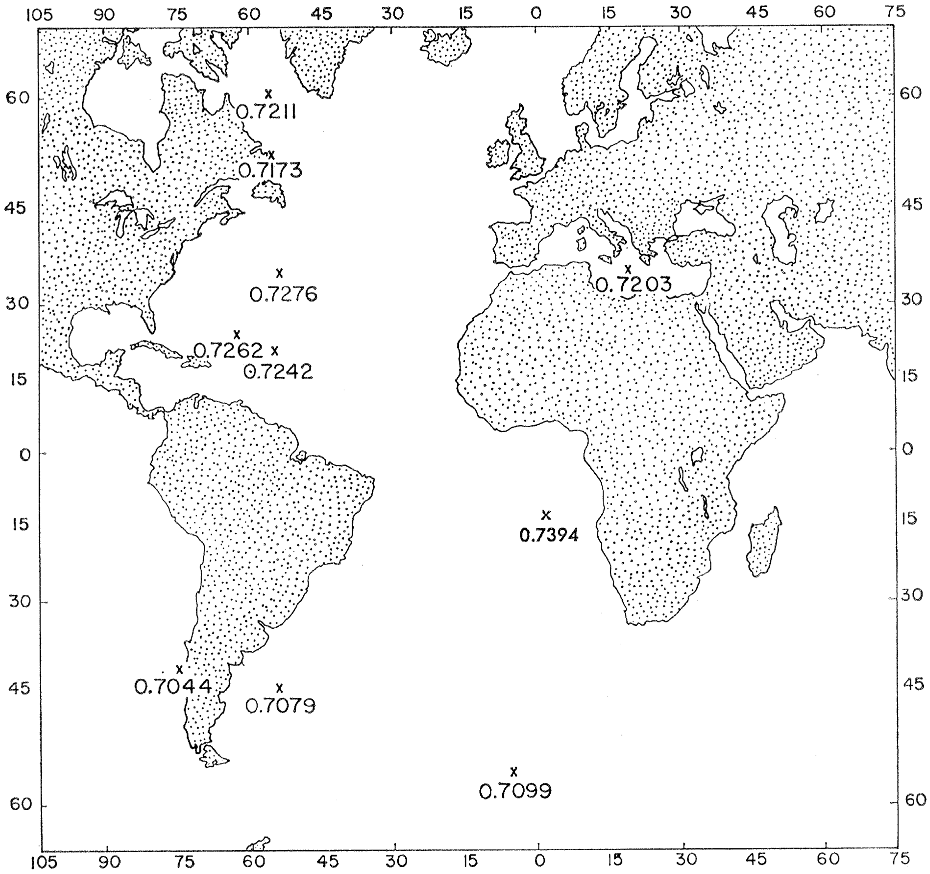

Fig. 1. Sources of samples and their Sr^{87}:Sr^{86} ratios from the detrital, aluminosilicate fraction of core-top sediments.

It is evident that strontium in detrital deep-sea sediments has not equilibrated isotopically with sea-water strontium. Only one sample (V14-60) has a strontium-isotopic composition indistinguishable from that of strontium in sea water. The ratios range from 0.7044 to 0.7394, a variation of 5 percent. A sample from off the coast of Chile (V17-55) has a ratio of 0.7044, which is near the low values (0.702 to 0.704) measured from oceanic volcanic rocks (*8*). Sample V12-73, from off the mouth of the Congo River, has a ratio of 0.7394, which may be a near-maximum ratio for deep-sea sediments. A similar high value has been reported for Lake Superior sediment, derived mainly from a 2.5-billion-year terrain (*9*). Because of the larger proportion of authigenic minerals and the generally lesser age of circum-Pacific rocks, sediments from the Pacific basin probably have lower $Sr^{87} : Sr^{86}$ values and a narrower strontium-isotope range than Atlantic sediments. The Argentine Basin sample (V17-121), the sample from off the mouth of the Congo River, and some of the samples from the North Atlantic have roughly equal amounts of illite. Thus, although illite may be the most important carrier of radiogenic strontium to the deep sea, strontium-isotope ratios do not correlate directly with illite abundances. Provenance appears

Table 2. Carbonate-free deep-sea sediments: strontium abundance and strontium-isotope composition. Clay mineralogy [weighted x-ray peak-area percentages, $< 2\text{-}\mu$ fraction *3*)]: M, montmorillonite; I, illite, K, kaolinite; C, chlorite. Strontium determined on silicate fraction by emission spectrography. S, single rhenium filament; D, double rhenium filament.

Core		Water depth (m)	Carbonate (%)	Clay mineralogy				Sr (ppm)	Ion source	Sr^{88}:Sr^{86} (meas.)	Sr^{87}:Sr^{86} (corr.)
No.	Source			M	I	K	C				
V17 -196	60°44′N, 57°50′W	2818	18.8	20	60	6	14	240	S	8.462	0.7211
V17 -199	53°42′N, 55°36′W	143	1.6	(sand)				530	D	8.286	.7173
A153-143	32°41′N, 50°54′W	5210	24.7	4	70	11	15	170	D	8.368	.7276
V16 - 15	25°15′N, 62°31′W	3225	3.5	10	69	10	11	190	S	8.403	.7262
V16 - 19	19°04′N, 53°46′W	4815	13.1	29	45	20	6	110	D	8.389	.7242
V17 - 55	40°33′S, 75°10′W	3636	2.7	56	21	3	20	280	D	8.316	.7044
V17 -121	43°58′S, 52°09′W	5784	2.4	30	43	5	22	250	S	8.338	.7079
V14 - 60	55°08′S, 04°57′W	3186	9.1	16 (diatomaceous)	56	6	22	20	D	8.306	.7099
V14 -135	35°42′N, 18°40′E	4027	11.3	27	50	12	11	200	D	8.317	.7203
V12 - 73	05°54′S, 09°53′E	3054	3.8	4	47	47	2	60	D	8.344	.7394

to be the chief factor influencing the strontium-isotopic composition of the noncarbonate fraction of deep-sea sediments.

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10. Supported by grants from the Esso Educational Foundation and the Westinghouse Educational Foundation for construction of the mass spectrometer; sponsored by NSF grant GP-2456. We thank Maurice Ewing, director of Lamont Geological Observatory, for the cores. The double-filament ion source was loaned by P. W. Gast (Lamont), whom we thank. J. A. Taylor assisted in the laboratory. Discussions with P. W. Gast and with V. R. Murthy (Univ. of Minnesota) were helpful.

4 April 1966

Optical Environment in Gemini Space Flights

In the debriefings of all astronauts of the Gemini and Mercury space flights there has been a continual insistence by the astronauts that "stars cannot be seen in the daytime," the only qualification of this statement being that planets and the moon or perhaps the brightest stars (for example, Sirius) could be seen.

If surface brightness of the daytime sky at orbital altitude precluded observation of stars, a real scientific dilemma would have to be faced. In this note we propose to (i) outline the scientific problem that would be presented if stars of second magnitude could not be seen in the daytime sky; (ii) consider the possible sources of background illumination which might account for the "nonobservation" of daytime stars by the astronauts; and (iii) present evidence from the on-board tape recorder used in Gemini V during the performance of the S-1 experiment which indicates that, if proper precautions are taken, first- and second-magnitude stars can be seen in the daytime sky.

The problem of observing stars against a background illumination was investigated theoretically and experimentally by Tousey (*1*). Figure 1, derived from Tousey's results, shows surface brightness of the sky as a function of the magnitude of stars that may be observed against this background. In the visible region of the spectrum, first-magnitude stars can be distinguished when background illumination is approximately 10^{-8} of the sun's surface brightness. If we suppose that astronauts are unable to see first-magnitude stars in the daytime, it must be assumed that the reason is a background illumination of the order of 10^{-8} ssb. (Hereafter, ssb will be used as a measure of the surface brightness in terms of the average sun's surface brightness.) For comparison we quote a few values in Table 1.

Table 1. Surface brightness at 5400 Å.

Entry	Object	Surface brightness (ssb)
A	Full moon	2×10^{-6}
B	Daytime zenith sky at sea level (-3 magnitude stars visible)	6×10^{-7}
C	Background against which first-magnitude stars may just be seen	10^{-8}
D	Zenith sky (night) with full moon (5.0 magnitude stars visible)	2×10^{-12}
E	Average nighttime sky away from Milky Way	4×10^{-14}

The conclusion then is that the required surface brightness of the background to limit observations to first-magnitude stars is a brightness of about 2×10^5 greater than the sea level nighttime sky (without a moon) and about 100 times dimmer than the daytime zenith sky at sea level.

We have divided the possible sources of illumination into three classes: (i) true illumination of the distant sky; (ii) a cloud around the spacecraft (spacecraft corona); and (iii) illumination on the spacecraft window.

True illumination of the distant sky could come either from scattering by air and dust above the spacecraft or from day airglow. Pressure at the spacecraft altitude is not greater than 10^{-9} of pressure at sea level so the expected scattered light should be about 10^{-15} ssb. This is 10^7 times too dim to account for the required 10^{-8} ssb (see entries B and C in Table 1).

About one-half of the nighttime sky brightness is produced by airglow. If we call this brightness 10^{-13} ssb, to enhance it to the required level would mean that the daytime airglow would have to be more than 10^5 times brighter than the nighttime airglow. In addition, this airglow would have to arise at altitudes in excess of 150 km. [Most of the nighttime airglow is in the layer centered at 90 km and is viewed in profile from above by the astronauts (*2*) at night.] The bright airglow postulated here would have to come from higher altitudes where the most pronounced airglow emission is the 6300 line of atomic oxygen. Barbier (*3*) has found evidence that the 6300-Å OI emission does indeed increase toward the sun (that is, as the elongation angle is decreased the airglow brightens). However, the absolute value of the 6300-Å brightness at 15° elongation is about 700 rayleighs. This corresponds to a continuum brightness of about 1 rayleigh per angstrom or about 10^{-13} ssb. If the background brightness is due to the 6300-Å line of OI, it would have to be enhanced by a factor of 10^5 from its value at 15° elongation angle as observed at night. It seems therefore that the required brightness of 10^{-8} ssb to just allow the observations of first-magnitude stars in the day-

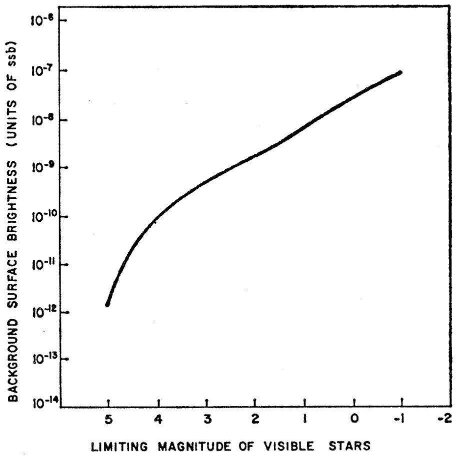

Fig. 1. The limiting magnitude of visible stars which may be seen in the presence of background illumination as a function of the brightness of the background.