

CHEMISTRY AND CLAY MINERALOGY OF CORES 8, 9, 10, NEW HARBOR, ANTARCTICA

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Abstract: Electrical conductivity and ionic composition of extracted solutions from ice-cemented permafrost from cores 8, 9, and 10 show that most of the sediments were deposited in a marine environment and that aggradation of permafrost during exposure of the sediments to subaerial conditions caused ionic concentration. Influx of brines capable of moving in permafrost is also suggested. Regions of low conductivity are interpreted either as freshwater episodes or textural discontinuities. Clay mineralogy can be separated into three major assemblages corresponding to the three major lithological units. The clay minerals show little weathering and complex irregular interstratification of mica-vermiculite-montmorillonite and chlorite.

1. Introduction

Because most of the world's ice is contained in the Antarctic, the present and past size and extent of the Antarctic ice sheets are of interest from several different viewpoints, including those that encompass problems of climatic change and prediction. The Dry Valley Drilling Project (DVDP) was organized as an international consortium to reconstruct the geological and geochemical history of the ice-free areas of southern Victoria Land and other regions of McMurdo Sound, Antarctica. This report presents and examines the chemical parameters of water extracts from, and the clay mineralogy of, cores 8, 9, and 10, which were obtained in the New Harbor area of southern Victoria Land (Fig. 1).

Chemical parameters such as electrical conductivity and ionic composition of solutions extracted from ice-cemented permafrost cores have been used to inter-

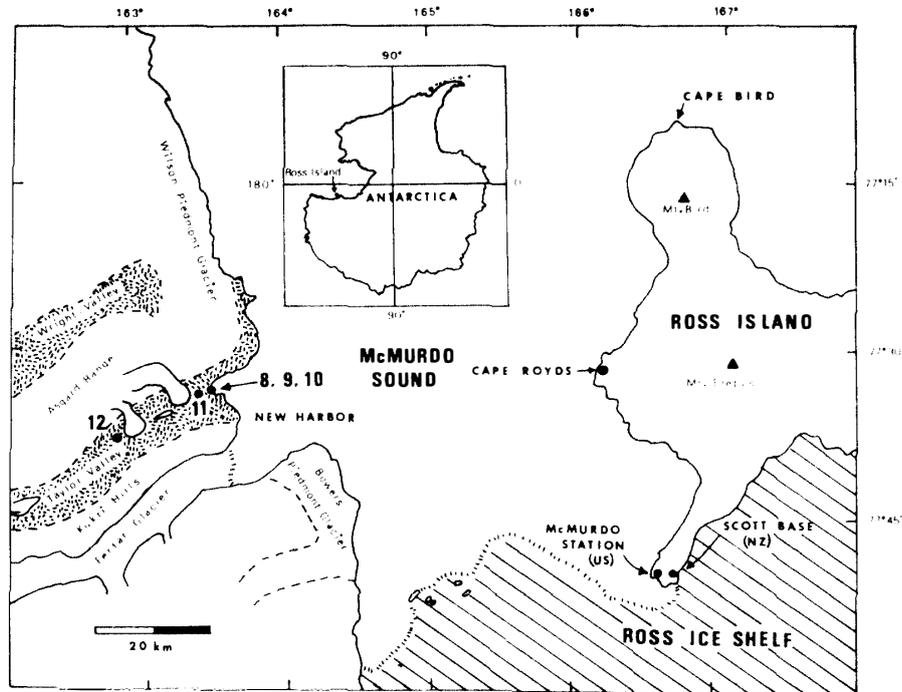


Fig. 1. The McMurdo Sound region, southern Victoria Land, Antarctica, and the New Harbor site of cores 8, 9, and 10.

pret the post-emergence history of marine sediments (O'SULLIVAN, 1966; BROWN, 1966; SELLMANN and BROWN, 1973) in the Barrow area, Alaska, and the Quaternary stratigraphy of terrestrial deposits near Fairbanks (PÉWÉ and SELLMANN, 1973; SELLMANN, 1967, 1968). PAGE (1978) and ISKANDAR *et al.* (1978) have more recently examined the salinity and chemical composition of pore water in permafrost in the offshore areas near Prudhoe Bay, Beaufort Sea. MCGINNIS *et al.* (1979) measured salinity and electrical resistivity of sections of cores from DVDP boreholes 10, 11, and 12, Taylor Valley, Antarctica, and used these data to reconstruct glacial events.

The mineralogy of the clay fraction from selected core sections was determined both to provide additional information on the lithology of the cores and to shed light on the provenance, weathering, and possible diagenesis of the sediments. WEAVER (1958) pointed out that clay minerals constitute more than 50% of the minerals in sedimentary rocks and that more than 95% of all sedimentary rocks contain clay minerals. This statement does not appear controversial, but disagreement exists whether the clay minerals in sediments, specifically in marine sediments, are detrital or diagenetic. MILLOT (1970) concluded that marine clay minerals are mostly detrital, but that some of them are subject to diagenesis after sedimentation. It would appear, therefore, that a systematic analysis of the clay minerals in the cores could provide information on the provenance of the sediments and on the environ-

mental conditions during diagenesis.

This report demonstrates that, in practice, a single interpretation of a particular chemical or mineralogical parameter can be misleading or erroneous. An awareness of this difficulty is implied in the work of O'SULLIVAN (1966), who noted that the salinity of sediments in a permafrost region is influenced by many factors, including texture, depth, ice content, and the history of the site. The environment of the sediments of cores 8, 9, and 10 has varied markedly—prior to, during, and subsequent to deposition. In principle, then, several different physical processes, acting singly, simultaneously, and/or at different times, could have left their fingerprints on the sediments and pore fluid. Consequently, some of the variations we have measured could have been caused in a number of different ways, and there is often nothing in the individual parameters themselves to allow an unambiguous historical interpretation. We have met this problem in two ways: (1) by identifying those physical processes which are active, or are thought to be active, in sedimentary sequences deposited in polar regions, and (2) by comparing different parameters, including those reported in the literature, in the hope that the comparison will allow a more complete and accurate interpretation.

In interpreting the chemical parameters of pore fluids, one must remember that both the water and the solutes could be: (1) connate, (2) unconconnate, or (3) some mixture of the two. There are three processes which might act, singly or in concert, to change the composition of a connate pore solution: (1) interaction with sediments, (2) freezing, and (3) ground-water flux.

Much interaction between sediment and solutes in a pore fluid is a function of the cation exchange capacity and weathering of the minerals in the sediment. Sediments in the clay and silt size fractions, often containing minerals having a high exchange capacity, tend to retain the solutes deposited with or passing through them, even under hydrodynamic conditions that might 'flush' a coarser sediment. This kind of effect is a possible explanation for the observations of O'SULLIVAN (1966), who reported a striking correlation between texture and salinity: silts displaying salinities 3 to 10 times, and clays 5 to 30 times, greater than that of adjacent sandy gravels. NAKAO *et al.* (1978) reported data for DVDP 14 which show a similar correlation. The retention of solutes could have a further effect. Several workers have suggested that ions held in clays and silts can act as a barrier to the passage of additional charged solutes. Under such conditions, the ions dissolved in ground water that passes through fine-grained material are filtered out and retained on the "upstream" side of the "charge barrier" (KHARAKA and BERRY, 1973; OLSEN, 1969).

Weathering of the sediments could be another cause of changes in the composition of the pore fluid. The presence of unfrozen water in frozen soils and sediments makes weathering and ionic migration a plausible phenomenon (ANDERSON, 1967, 1968, 1973; UGOLINI and ANDERSON, 1973; MURRMANN, 1973). FRIPIAT

et al. (1965) have shown that the dissociation of water in films next to the surface of silicates is many times greater than in bulk. The high ionic strength of the film of unfrozen water surrounding the mineral grains leads to corrosion of the silicate surface. Ions are thus liberated from the solid phase and become available to move in the unfrozen water. MURRMANN and HOEKSTRA (1970) and MURRMANN *et al.* (1968) have shown that although ion diffusion is considerably lower in frozen soils than in unfrozen ones, the diffusion coefficients in the temperature range of 0°C to -15°C are relatively high, on the order of $1 \times 10^{-7} \text{ cm}^2 \text{ s}^{-1}$ for silt soil at -3°C. When the soil is water saturated and in a closed system, conditions existing in permafrost, the diffusion coefficient of an ion is virtually independent of total water content and thermal gradients, but determined by the quantity of unfrozen water, which is controlled by temperature (MURRMANN and HOEKSTRA, 1970; MURRMANN *et al.*, 1968; HOEKSTRA, 1965).

The effect of freezing on pore fluids are complex. It has long been known that dissolved ions are almost entirely excluded from the ice phase as freezing progresses (THOMPSON and NELSON, 1956; TERWILLIGER and DIZIO, 1970; HALLET, 1976; ANISIMOVA, 1978; YAS'KO, 1978). HALLET *et al.* (1978) has recently shown that cryogenic precipitates may originate through rejection of salts during freezing of ionic solutions. In principle, then, the freezing of an originally brackish pore fluid could result in the formation of two adjacent and physically separate 'strata' of solutions, one fresh and frozen, the other, expelled from its original position, highly saline and liquid. In fact, the geometry and rate of advance of the freezing front must have a significant effect on the degree of solute segregation found in a permafrost section. As implied above, material type must also play a significant role in determining how much solute is held in place during freezing. The freezing of pore fluids not only is responsible for ionic exclusion, but also tends to segregate the ions as function of the lowering of the temperature. The early experiments of THOMPSON and NELSON (1956) stood to demonstrate that chemical evolution of sea water is possible when it is subjected to progressive freezing. Following a sequential segregation of salts, the composition of sea water changes and evolves. At a temperature of -22.9°C, after calcium carbonate, mirabilite, and sodium chloride dihydrate have crystallized, the original sea water has become a brine enriched in magnesium, potassium, and calcium chloride (THOMPSON and NELSON, 1956). As we will later compare our data to those of STUIVER *et al.* (1976), we also note that hydrogen and oxygen isotopes in water are fractionated between the liquid and solid phases during freezing.

Hydrodynamic systems are obvious mechanisms for the redistribution of water and solutes; they can in principle produce very complex effects. For example, connate solutions in bedded sediments might be preferentially removed from the coarser, more permeable layers, owing both to larger ground-water fluxes in those layers and to sediment properties of the kind discussed above. The situation can be further complicated in a permafrost region by the presence of frozen ground

which acts as a zone of very low permeability. Prior to the aggradation of permafrost, coarse materials are most likely to be affected by extensive ground water motion. Subsequent to aggradation, motion can occur only in fluids saline enough to remain unfrozen. This factor favors a reversal of hydrogeologic roles peculiar to permafrost terrains: namely, that fluids are likely to move most rapidly (albeit very slowly) in fine-grained materials, precisely because these materials are most likely to retain saline fluids and unfrozen water (ANDERSON, 1967, 1968). A further hydrogeologic complexity is introduced by the increase in volume that occurs during the phase change from liquid water to ice. Even in the absence of the forces that "normally" give rise to ground water flow, some flow must occur in a freezing sediment in order to accommodate the expansion. Some of these phenomena are discussed by CARTWRIGHT and HARRIS (1978), who describe modern hydrogeologic systems in the Dry Valleys.

2. Materials and Methods

Cores 8, 9, and 10 were obtained at the eastern end of Taylor Valley, southern Victoria Land, Antarctica, at an elevation of 1.9 m, and about 50 m from the southwest shore of New Harbor (MCKELVEY, 1975). The cores penetrated the same stratigraphic units and are therefore considered identical. Core 8 was drilled to a depth of 157 m; core 9 was obtained to secure a more complete record of the upper 38 m of core 8. The three cores encompass a total length of 185.47 m. Measurements of chemical parameters were made on 1:5 sediment-water extracts. 10 g of air-dried sediments were mixed with 50 ml of water; the mixture was agitated, allowed to stand for 10 minutes, and centrifuged. The conductivity of the decantate was measured with a Yellow Spring Model 31 conductivity bridge. Chloride and SO_4^- were determined by autoanalysis (Technicon Industrial method AA11, 99-70W, 1971; and AA11, 226-72W, 1971, respectively). Ca^{2+} , Mg^{2+} , Na^+ , and K^+ were determined by atomic-absorption spectrophotometry (Instrumentation Laboratory 353).

The clay fraction ($<2 \mu\text{m}$) was obtained from selected samples after salt and iron removal (MEHRA and JACKSON, 1960). The deferrated clay fraction was saturated with Mg^{2+} and glycolated. A clay aliquot was saturated with K^+ and heated at 300°C and 500°C . X-ray diffractograms were obtained with a Picker unit with $\text{CuK}\alpha$ radiation and a Ni filter; the unit was operated at 30 kV and 20 mA. A separate aliquot of clay was treated with dimethylsulfoxide (DMSO) to differentiate between kaolinite and chlorite (ABDEL-KADER *et al.*, 1978).

3. Results and Discussion

In examining the electrical conductivity values in Table 1 and Fig. 2, major and

Table 1. Electrical conductivity (EC) for DVDP cores 8, 9, and 10, New Harbor, Antarctica.

Hole 8		Hole 9		Hole 10	
Depth (m)	EC 25°C μ mhos/cm	Depth (m)	EC 25°C μ mhos/cm	Depth (m)	EC 25°C μ mhos/cm
		6	1800		
		9	1800		
		11	1200		
		19	1000		
		21	2700		
		25	1900		
		27	900		
		35	800		
44	890			37	800
46	3100			46	2500
49	960				
52	940				
54	1700				
57	4400				
58	4600				
60	5700				
63	4300				
65	3900				
68	5600			68	1800
71	2600				
74	2300				
78	2400			83	1900
90	3200			87	1200
98	2200				
103	3500			109	2200
112	3000			112	2900
113	2300			113	2500
118	1400			119	1100
123	2500				
123	1900				
126	4800				
131	2500				
133	3500			133	3500
136	1900				
146	2900				
146	2800				
147.9	2500				
148	8400				
150	6400				
154	3400				
155	5500			155	5500
156	7100			158	10000
				168	7900
				171	900
				175	3700
				180	900
				183	8200

minor fluctuations can be observed. The major fluctuations occur in the lower part of the core, from 140 m to 183 m, and in the upper part, from 6 m to 71 m. The central part of the core, from 71 m to 148 m, shows minor fluctuations. Because conductivity was measured on a 1:5 extract, the values obtained are internally

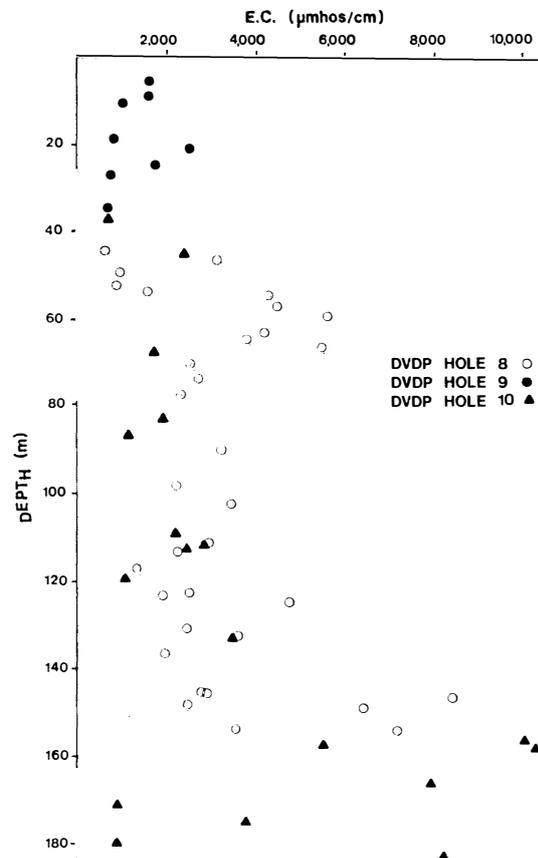


Fig. 2. Electrical conductivity values for cores 8, 9, and 10 (New Harbor) plotted according to depth (meters).

consistent but do not directly measure the original concentration of ions in the interstitial water. However, the fluctuations we measured, except for some discrepancies, parallel rather closely the fluctuations in the salinity curve that MCGINNIS *et al.* (1979) obtained from fluids squeezed from the sediments.

From 148 m down to 183 m, there are widely varying conductivities, with maxima at 148 m, 156 m, 158 m, 168 m, and 183 m. Lower conductivity values are recorded at 171 m and 180 m (Table 1). The salinity curve of MCGINNIS *et al.* (1979) also shows considerable fluctuation below 140 m, which they interpreted as evidence of alternating periods of exposure to freshwater and saltwater environments. They supported the view that the pore water is connate and that the salt enrichment was due to permafrost aggradation and segregation of brines at the base of the permafrost. Low salinity values were explained, alternatively, by (1) sea level lowering, (2) crustal uplift, (3) impounding of freshwater by advances of the Ross Ice Shelf in New Harbor, or (4) a combination of the foregoing. We, too, ascribe the high conductivity values at core depths of 148 m, 150 m, 168 m, and 183 m to brine concentration in the fine-grained marine sediments during freezing. The relatively high values for Na, K, Ca, Mg, and SO_4^- indicate that repeated freezing

Table 2. Chemical characteristics of selected samples for DVDP cores 8, 9, and 10, New Harbor, Antarctica.

Core	Depth (m)	Ion content, ppm in 1:5 sediment-water extract					
		Na ⁺	K ⁺	Ca ²⁺	Mg ²⁺	Cl ⁻	SO ₄ ²⁻
9	21	450	30	21	20	724	25
9	25	380	24	9	12	553	28
10	37	150	14	4	6	250	25
10	46	575	38	5	8	700	155
8	57	970	89	10	15	1426	37
8	60	1170	93	12	25	1812	53
10	68	150	33	4	4	250	45
10	83	350	30	6	11	300	35
10	87	175	21	6	7	450	120
10	109	475	31	10	10	700	30
10	112	640	48	13	15	900	135
10	113	520	54	9	11	1100	140
10	119	225	25	9	5	300	50
10	133	600	47	17	42	1175	135
8	148	520	32	16	10	733	35
10	158	2100	99	85	163	3700	107
10	171	170	10	11	9	278	11
10	180	180	11	4	6	251	9
10	183	1900	84	54	79	2980	86

and ionic segregation may be involved (Table 2). Although the low conductivity value of 900 μ mhos at 171 m coincides with a very coarse sand unit, the low value at 180 m occurs in fine-grained sediments (POWELL and BARRET, 1975; POWELL, 1978). Consequently, we infer that the coarse sands at 171 m failed to retain the ions during the aggradation of permafrost. On the other hand, the low conductivity of the fine sediments (64% finer than 8 ϕ , POWELL, 1978) at 180 m depth must be explained by a freshwater event.

The permafrost oxygen-isotope ratios (STUIVER *et al.*, 1976) indicate freshwater values between 150 m and 180 m. However, only at depths 171 m and 180 m could conductivity values indicate freshwater conditions and thus coincide with the isotopic ratios. The high salinities above 171 m and 180 m, ascribed in this report to marine conditions, do not correlate with the oxygen-isotope ratios. Perhaps the salts are connate, but the original interstitial fluids were replaced by freshwater.

Between 68 m and 146 m the conductivities were relatively low, the average value being approximately 2400 μ mhos/cm. The electrical conductivity curve

between these two depths displays a number of single peaks and valleys indicating fluctuations in the concentration of dissolved ions. Moderate conductivity values on the order of 4800 and 3000 $\mu\text{mhos/cm}$ occur, respectively, at 126 m and 90 m. Low values, 1100 and 1200 $\mu\text{mhos/cm}$, occur at 119 and 87 m. Moderate values occur at 146 m (2800 $\mu\text{mhos/cm}$) and 68 m (1800 $\mu\text{mhos/cm}$). The high and low conductivity values are consistent with the salinity distribution reported by MCGINNIS *et al.* (1979). We interpret this unit as reworked marine sediments (WEBB and WRENN, 1979; BRADY, 1979) which have been affected by freshwater. It is not possible to precisely determine the number of cycles of marine and freshwater environments from our electrical conductivity values. However, values at the 119 m and 87 m depths seem to indicate terrestrial environments. Between core depths of 146 m and 68 m, BEGET (1977) distinguished five diamictons, two of which displayed a strong microfabric. The lowest was from 114 m to 120 m and the upper from 74 m to 68 m. He ascribed the lowest to a basal till recording the incursion of the Ross Sea Ice. This interpretation agrees with freshwater isotopic ratios reported by STUIVER *et al.* (1976), and with our low conductivity values. The upper region (74 m to 68 m), which shows a strong microfabric, failed to show freshwater isotopic values but correlated well with the low salinity of MCGINNIS *et al.* (1979) and with our low conductivity values. The uppermost unit above 71 m has high initial conductivity values, which decrease rather uniformly to values of 1000–2000 $\mu\text{mhos/cm}$. Oxygen isotopes (STUIVER *et al.*, 1976) and microfabric data (BEGET, 1977) show this to be a marine section. From the surface to a depth of 24.86 m BRADY (1978) reported both marine and nonmarine diatoms. According to WEBB and WRENN (1979), the lowermost part is a diamicton, possibly deposited by ice rafting, and the upper section is deltaic. A marine delta is sub-aerially exposed during growth and is affected by both influx of freshwater and by permafrost aggradation. Combined with the influence of textural variability, these effects may explain the conductivity measurements in this interval. The lowest conductivity recorded coincides with the unit with the largest clasts, and the highest conductivity with fine-grained diamictons (POWELL, 1978).

3.1. Ionic concentration and ionic ratios

Ionic concentration in the 1:5 sediment-water extract shows that Na and Cl are the predominant ions all throughout the cores 8, 9, and 10 (Table 2). When Na, K, Ca, and Mg are plotted in the diagram of TORII *et al.* (1978), the points for cores 8, 9, and 10 tend to concentrate in the left corner of the diagram, close to sea water composition (Fig. 3). The points for Taylor Valley and Cape Royds soils also appear to betray the marine influence. The points for the Victoria Valley soils (VV I through VV IV) tend to approach glacier melt water and the Onyx River composition (Fig. 3). Although the 1:5 sediment-water extract tends to dilute the original interstitial fluids and, thus, favor the solution of Na and K, the chemical

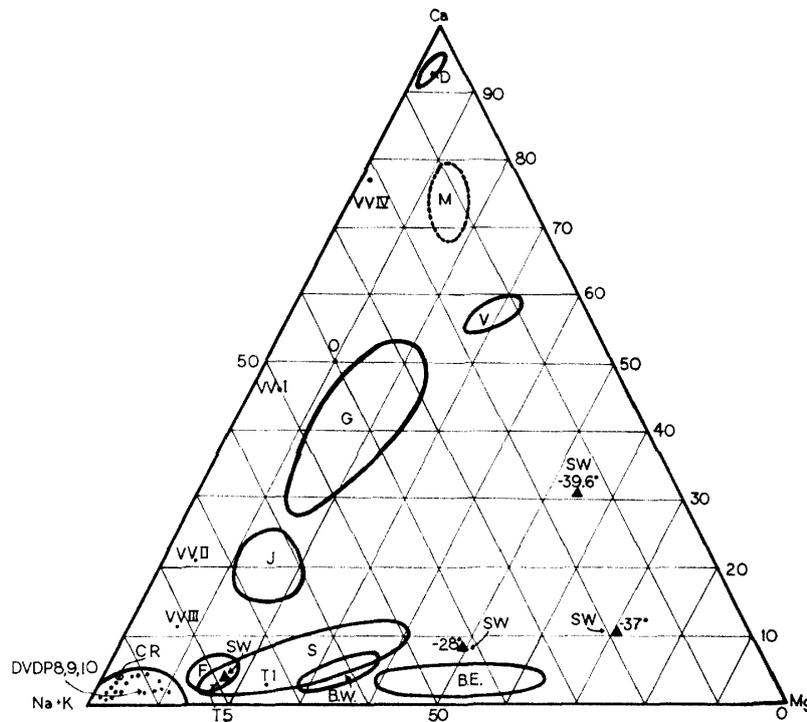


Fig. 3. Chemical composition of snow and ice, glacier melt waters, lakes, sea water below freezing, soil-water extracts, sediment-water extracts of cores 8, 9, and 10. Dry Valley, Antarctica. Legend: S=snow and ice; M=Miers Lake; J=Joyce Lake; F=Fryxell Lake; BW=Lake Bonney West; BE=Lake Bonney East; V=Lake Vanda; D=Don Juan Pond; VVI to VVIX=soil-water extract, Victoria Valley; TI=soil-water extract, Taylor Valley Site 1; T5=soil-water extract, Taylor Valley Site 5; CR=soil-water extract, Cape Royds; DVDP 8, 9, and 10=sediment-water extract, cores 8, 9, and 10 to New Harbor; ▲=sea water below freezing (Modified from T. TORII *et al.*, 1978).

characteristics of the extract seem to indicate sea water contamination for all the core depths. A marine origin of the core unit below 150 m is suggested by the microfauna analysis of WEBB and WRENN (1979). These authors recognized an unconformity at 154 m, and they ascribed the sediments and microfauna below this depth (Zone 11) to marine deposition in a sublittoral fjord environment extending from early Pliocene to late Miocene. BRADY (1978), on the basis of diatoms in core 10, recognized an early Pliocene marine unit from 137 m to 183 m. Additional evidence of marine conditions is provided by BEGET (1977) and PORTER and BEGET (1978), who believed that the sediments between 148 m and 183 m were initially deposited in a deep-water marine environment, as indicated by rare drop stones and graded beds resembling turbidities.

The ionic profile for the next unit, 150 m to 68 m, still betrays a marine influence

Table 3. Ionic ratios for selected depths, DVDP cores 8, 9, and 10, New Harbor, Antarctica.

Core	Depth (m)	Na/Cl	Na/Ca	Na/K	Ca/Cl	Cl/SO ₄	Ca/Mg
8	21	0.62	21.4	15.0	0.03	28.9	1.05
9	25	0.69	42.2	15.8	0.02	19.7	0.75
10	37	0.60	37.5	10.7	0.02	10.0	0.67
10	46	0.82	115.0	15.1	0.01	4.5	0.63
8	57	0.68	97.0	10.9	0.01	38.5	0.67
8	60	0.65	97.5	12.6	0.01	34.2	0.48
10	68	0.60	37.5	4.5	0.02	5.5	1.00
10	83	1.17	58.3	11.7	0.02	8.6	0.55
10	87	0.39	29.2	8.3	0.01	3.7	0.86
10	109	0.68	47.5	15.3	0.01	23.3	1.00
10	112	0.71	49.2	13.3	0.10	6.6	0.87
10	113	0.47	57.8	9.6	0.01	7.8	0.82
10	119	0.75	25.0	9.0	0.03	6.0	1.80
10	133	0.51	35.3	12.8	0.01	8.7	0.40
8	148	0.71	32.5	16.3	0.02	20.9	1.60
10	158	0.57	24.7	21.2	0.02	34.6	0.50
10	171	0.61	15.5	17.0	0.04	25.3	1.20
10	180	0.72	45.0	16.4	0.02	27.9	0.67
10	183	0.64	35.2	22.6	0.02	34.7	0.68
Sea water		0.53	26.0	28.0	0.02	7.20	0.30

(Table 2 and Fig. 3).

The uppermost unit shows a marine influence (Table 2); however, at a depth of 37 m, a low electrical conductivity is recorded. The Na and Cl concentrations are also low, perhaps indicating a terrestrial environment. Especially significant is the Na/Cl ratio, which in core 10 slightly exceeds that of normal sea water (Tables 3 and 4), suggesting that originally the interstitial water was sea water.

Na/K ratios for the sediments of core 10 range from 4.5 to 22.6, compared with a ratio of 28.0 for sea water. Potassium enrichment, resulting in a lower ratio, may be due to weathering of K-rich minerals, or it may be the result of the effect of temperature on the exchangeable cations. With increasing temperature of the sediments, K, Na, and Li are desorbed and diffused into the interstitial water (PAGE 1978). As all the samples were treated alike, any artificial temperature effect should have been the same in all samples. An examination of the clay mineralogy (Table 5) at depths where Na/K ratios are low failed to reveal an unusual degree of weathering of either the mica or feldspar. Finally, the K enrichment could have been caused by influx of freshwater from land when the sediments were exposed to terrestrial

Table 4. Ionic ratios for lakes, soils, and DVDP cores 8, 9, and 10, McMurdo region, Antarctica.

	Na/Cl	Na/Ca	Na/K	Ca/Cl	Cl/SO ₄	Ca/Mg
<i>Lakes</i>						
Vanda 50 m (O)	0.27	0.36	3.90	0.76	17.0	3.90
Vanda 64 m (O)	0.09	0.23	7.30	0.38	90.0	3.4
Don Juan (O)	0.06	0.09	73.0	0.50	13.0	73.0
N. Fork Saline Pond (O)	0.64	8.20	32.0	0.08	2.10	0.30
<i>Soils</i>						
Taylor Valley Site 1 (L)	0.45	35.0	6.1	0.014	34.9	0.09
Taylor Valley Site 5 (L)	0.54	33.3	16.8	0.016	20.4	0.13
Ross Island (T.U.)	0.46	10.34	2.41	0.04	6.36	4.3
Beacon Valley (T.U.)	1.70		11.93		0.09	
Ross Island (T.U.)	0.60	14.2	1.72	0.043	5.20	3.3
<i>Cores</i>						
Cores DVDP 8, 9, and 10 (range)	0.62–1.17	15.0–115.0	4.5–22.6	0.01–0.03	34.7–3.75	0.4–1.8
Sea water	0.53	26.0	28.0	0.02	7.20	0.30

(O): OBERTS, G. L. (1973): The chemistry and hydrogeology of Dry Lakes, Antarctica. M. S. Thesis, Dept. Geology, N. Illinois Univ., 55 p.

(L): LINKLETTER, G. O. (1970): Weathering and soil formation in Antarctic Dry Valleys. Ph. D. Thesis, Univ. of Wash., 122 p.

(T. U.): TEDROW, J. C. F. and UGOLINI, F. C. (1966): Antarctic Soils. Antarctic Soils and Soil Forming Processes, ed. by J. C. F. TEDROW. Washington. Am. Geophys. Union, 161–177 (Antarctic Research Series, 8).

conditions. Core samples at 68 m, 87 m, 113 m, and 119 m show Na/K ratios diverging from that of sea water, suggesting freshwater environments (Table 4). Calcium/magnesium ratios in core 10 show that calcium is enriched with respect to sea water (Tables 3 and 4). This enrichment of Ca may be due to either the cation exchange reactions or the displacement of Ca from the exchange sites due to the influx of K.

The Cl/SO₄⁻ ratios (Tables 3 and 4) vary from 3.75 to 34.7 and, in general, tend to be larger than for sea water (Table 4). The lower part of the core, from 148 m to 183 m, appears to be consistently depleted in SO₄⁻. This depletion may be due to reduction in the presence of Desulfovibrio bacteria. Enrichment in SO₄⁻ with respect to Cl⁻ may indicate leaching of Cl⁻ and Na⁺. The region between 114 m and 120 m in the core could evidence such a leaching episode. Both the Na/K ratio and the electrical conductivity are low at these depths.

3.2. Clay mineralogy

The clay mineralogy of core 10 consists of a complex association of interstratified

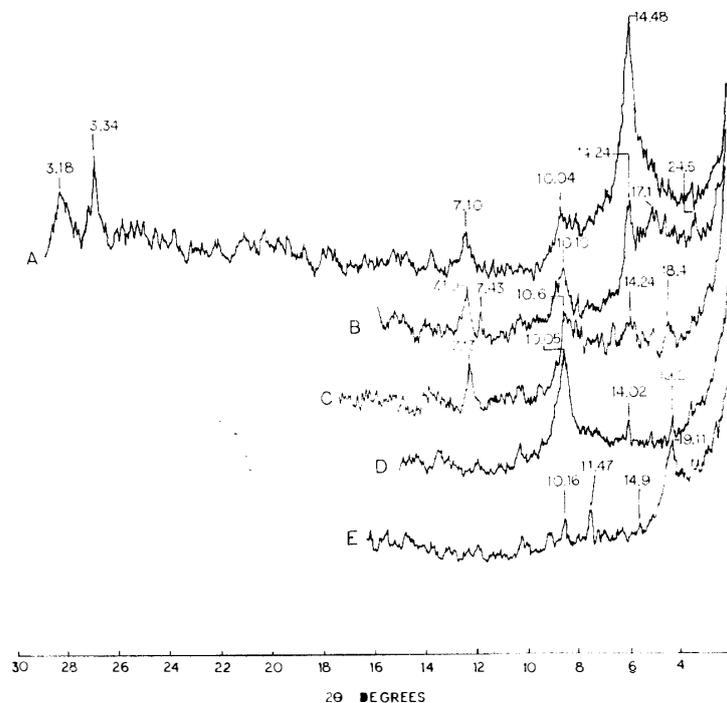


Fig. 4. X-ray diffractograms of clay fraction $< 2 \mu\text{m}$, oriented mount, $\text{CuK}\alpha$ radiation, core 10, 148 m depth.

A=Mg-saturated, B=Mg/glycol saturated, C=K-saturated ambient, D=K-saturated heated at 550°C , E=Li-DMSO treated.

montmorillonite-vermiculite-chlorite-mica. Most of the interstratification is irregular; however, regular interstratification leading to the formation of superlattices was also observed. The diffractograms in Fig. 4 are from the clay fraction of the 148 m depth. The diffractogram of Mg-saturated clay shows a prominent peak at 14.48 \AA and two smaller peaks at 10.04 \AA and 7.10 \AA . Upon glycolation the 14 \AA peak resolves itself into a number of diffraction intensities. A much reduced peak at 14.24 \AA indicates that a sizable portion of the 14.48 \AA (Mg-saturation) was montmorillonite, now expanded at high-order spacings. The numerous small peaks between 17.1 and 24.5 \AA reveal the presence of irregularly interstratified montmorillonite-vermiculite with some regular interstratification forming superlattices. The broad mica peak (10.10 \AA) indicates irregularly interstratified mica-vermiculite. The 7.10 peak may represent the second order diffraction of vermiculite as well as the presence of kaolinite. K-saturation and heating at 300°C and 500°C tend to collapse the expansible minerals. The collapsing of the interstratified montmorillonite-vermiculite induces the reinforcement of the 10 \AA peak. Treatment with dimethylsulfoxide (DMSO) (ABDEL-KADER *et al.*, 1978) clearly showed the presence of kaolinite at 11.4 \AA , mica at 10.16 \AA , and a peak for both montmorillonite and

vermiculite at 19.11 Å (Fig. 4).

In the lower part of the core, from 158 m to 182 m, the interstratification is more irregular and complex (Table 5). Diffractograms of Mg-saturated clay show little resolution of the individual minerals and a high degree of irregularities. Interstratified, highly variable montmorillonite-vermiculite is the dominant phase; however, non-interstratified vermiculite, mica and chlorite species, with some kaolinite, appear as minor and trace components together with quartz and feldspar. From 71 m to 146 m (Table 5) the core showed a weakly developed interstratification of montmorillonite-vermiculite-chlorite and mica but only a few individual components of these mixed-layer complexes were noted. Feldspars were abundant throughout the interval, except at 90 m and 71 m, where quartz prevailed.

The section of the core from 60 m to the surface (Table 5) showed a complex assemblage of interstratified phyllosilicates similar to that in the underlying unit but containing more unmixed individual components, such as vermiculite, mica, and kaolinite. This upper section was depleted in clay-size quartz and feldspars.

The clay mineral assemblages throughout the core indicate that the sediments

Table 5. Clay mineralogy by X-ray diffraction analysis for selected samples from core 10 (<2 μm fraction).

Depth (m)	Montmorillonite, vermiculite, chlorite, and mica interstratified	Vermiculite	Mica	Chlorite	Kaolinite	Quartz	Feldspar	Amphibole	Pyroxene
6	M					m-t	m-t		
21	M					m	m	m	m
25	M		t	t					
57	M	t	t		t				
60	M	t	t		t	m		m	
71	M					M	m		
90	M					M		m	
118	M					m	M	m	
120	M			t			M	m	
123	M						M		
133	M						M	m	
146	M						M		
148	M				t	m	m		
158	M	m	m	m-t	t	m	m		
171	M	m	m	m	t	M	M	m	
179	M	m	m	m-t		m	m	t	t
182	M	m	m	m-t	t	m	m	m	

M= major; m= minor; t= trace.

experienced a low level of weathering. Mixed layers are formed readily and represent either the initial stage in weathering (SAWHENY, 1977) or the mature weathering products of such regions as deserts, where little weathering occurs (SAWHENY, 1977; M. L. JACKSON, personal communication, 1978). The mica-vermiculite-montmorillonite interstratification indicates only a partial removal of K from freshly cleaved mica flakes. Removal of K results in structural changes and in reduction in layer charges, which in turn results in the appearance of mixed layers (NORRISH, 1973). Clay minerals identified in the soils of different regions of continental Antarctica are largely the products of weathering of mica (CLARIDGE, 1965) but show a more advanced stage of weathering than the minerals in core 10. Furthermore, in soils younger than 10000 years, montmorillonite is rather common as a separate phase in the fine clay ($<2 \mu\text{m}$), with the feldspar being restricted to the coarser fractions (UGOLINI and JACKSON, 1979). Weathering of feldspar into mica, and subsequent weathering of mica into montmorillonite, has been recognized in an old soil in the Wright Valley exposed to an intensive period of weathering approximately 4 million years ago (JACKSON *et al.*, 1977). PIPER and PE (1977) detected montmorillonite and iron-rich chlorite in Oligocene to Quaternary turbidities derived from Victoria Land. The mixed layer phyllosilicates observed in core 10 could also have formed by diagenetic sorption of K and Mg by the sediment in the marine environment. Different authors (in MILLOT, 1970) have postulated a diagenetic origin for some newly formed illite (mica) and chlorite. This type of transformation could be invoked to explain the mineralogy of lowermost unit (158 m–182 m) of core 10. Degraded mica and montmorillonite washed from the soils could, in the marine environment, have recaptured the K and Mg to reconstitute the minerals vermiculite, chlorite, and mica. The persistence of mixed layers could indicate that the transformation and regularization of the lattices were incomplete.

Alternatively, montmorillonite could have formed *in situ* in the permafrost as recently suggested by USKOV (1978). According to this author, hydromica, chlorite, and more slowly kaolinite are changed into montmorillonite in the Pleistocene permafrost on Yakutia, USSR. Because of heterogeneity of the sediments and the complex history of the cores analyzed, it would be difficult to support Uskov's contention to explain the observed clay assemblage in core 10.

4. Summary and Conclusions

In summary, the electrical conductivity values and the ionic ratios seem to indicate that most of the sediments in core 10 were deposited in a marine environment and that aggradation of permafrost during subaerial exposure of the sediments caused ionic concentration and segregation. Regions of low conductivity that display more "terrestrial-like" ionic ratios may be interpreted as evidence of either freshwater episodes or the exclusion of salt from coarse sediments during slow down-

ward movement of a freezing front.

Clay mineralogy for cores 8, 9, and 10 can be separated into three major assemblages. Below a core depth of 158 m there is a unit of uniform clay mineralogy. Although showing complex irregular interstratification, this unit displays both non-interstratified individual mineral species and traces of kaolinite, and is the most weathered unit in the core. According to BEGET (1977), the lithological assemblages for this unit indicate a constant source from eastern Taylor Valley.

Above 158 m and below approximately 60 m the clay mineralogy, although similar to that of the underlying unit, shows an abundance of feldspar and only a few individual mineral species. This is the least weathered unit. The lithology and texture of the sediments suggest a source southeast of Taylor Valley (BEGET, 1977).

The clay mineralogy in the uppermost unit has characteristics of both the lower two units. This is expectable inasmuch as the unit represents a delta that received sediments from the glacial drift reported in the middle unit, which extends up Taylor Valley. In addition, the delta received sediments from the eastern Taylor Valley material, the source of the lowermost core unit.

The clay mineralogy of the entire core does not show distinct weathering horizons or soil formation episodes. The predominance of mixed layer clay minerals and the presence of clay-size feldspars and amphibole suggests that the weathering environment in Antarctica has been, since the early Pliocene, similar to the present—a polar desert.

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