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THE INFLUENCE OF CLAY-SIZED PARTICLES ON SEISMIC VELOCITY FOR CANADIAN ARCTIC PERMAFROST

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ABSTRACT

Seismic wave velocities have been measured on 37 unconsolidated permafrost samples as a function of temperature in the range -16° to +5°C. The samples, taken from a number of locations in the Canadian Arctic Islands, the Beaufort Sea and the Mackenzie River, were tightly sealed immediately upon recovery in several layers of polyethylene film and maintained in their frozen state during storage, specimen preparation, and until they were tested under controlled-environmental conditions. During testing, the specimens were subjected to a constant hydrostatic confining stress of 0.35 MPa (50 psi) under drained conditions. At no stage was a deviatoric stress applied to the permafrost specimens. The fraction of clay-sized particles in the test specimens varied from almost zero to approximately 65 %. At temperatures below -2°C the compressional-wave velocity was observed to be a strong function of the fraction of clay-sized particles, but only a weak function of porosity. At temperatures above 0°C the compressional-wave velocity was observed to be a function only of porosity, with virtually no dependence upon the fraction of clay-sized particles. Calculation of the fractional ice content of the permafrost pore space from the Kuster and Toksöz theory showed that for a given fraction of clay-sized particles the ice content increases with an increase in porosity. It is concluded that the compressional-wave velocity for unconsolidated permafrost from the Canadian Arctic is a function of the water-filled porosity, irrespective of the original porosity, clay content, or temperature.

INTRODUCTION

Over the past 2 decades a considerable interest has resulted in the exploration for and exploitation of the large mineral resources to be found in the Far North. Some 20% of the earth's dry land is covered with permanently frozen ground, or permafrost, including large areas of northern Canada and Alaska. It is therefore necessary to study the effects of permafrost on the interpretation of geophysical surveys performed in these areas. Interpretation of seismic reflection data, an important technique in exploration geophysics, depends upon seismic velocity control, particularly in near-surface sediments. It is these sediments that often are in the permanently frozen state in the Far North.

Problems associated with the interpretation of seismic and other geophysical surveys conducted in areas where permafrost is present have been discussed by Scott et al. (1979). A number of studies have demonstrated that the presence of ice in the pore spaces of sediments can result in large increases in seismic velocity over the case for which the interstitial water is unfrozen. Besides ground temperature, other factors influencing the fraction of ice formed in the pore spaces of sediments are the moisture content, the pore sizes and shapes, the pore-water chemistry, and the states of stress in the sediment and pore water. Theoretical aspects of these points have been discussed in considerable detail in a review prepared by Anderson and Morgenstern (1973). Experimental studies of the factors referred to have been reported by Timur (1968), Nakano et al. (1972), Kurfurst (1976), King (1977), Pandit and King (1979), King et al. (1982), and King and Garg (1982).

From his studies of elastic-wave propagation in a number of water-saturated sedimentary rocks at permafrost temperatures, Timur (1968) concluded that as the temperature was reduced below 0°C in these rocks, ice formed first in the larger pore spaces and then in progressively smaller ones as the temperature was reduced still further. He ascribed this behavior to the progressively increasing interfacial forces asssociated with larger specific surface areas as the pore sizes became smaller, coupled also with the salinity of the interstitial water. Nakano et al. (1972), King (1977), Pandit and King (1979), and King et al. (1982) have observed behavior confirming Timur's hypothesis during measurements of elastic-wave velocities, complex electrical resistivity, and thermal conductivity on samples of natural permafrost and other water-saturated porous rocks at temperatures below 0°C.

Pandit and King (1979) have studied the effect of changes in pore-water salinity on the elastic-wave velocities and complex electrical resistivity of frozen, water-saturated samples of porous, sedimentary rocks. They concluded that an increase in salinity of the pore water decreased the ice content of the pore spaces at a given temperature below 0°C, with the decrease most noticeable at temperatures below, but close to, 0°C.

The results of a study to determine the effects of clay-sized particles on elastic wave velocities in unconsolidated, water-saturated permafrost, are presented in this paper. In earlier studies, elastic wave velocities were measured by the author under contracts with the Geological Survey of Canada [Kurfurst (1976)] and Panarctic Oils Ltd. [King et al. (1982)] on twenty permafrost samples from shallow boreholes (3 to 25 m deep) along the Mackenzie River and from sub-seabottom boreholes (7 to 81 m deep) in the Beaufort Sea. Results of tests on a further seventeen permafrost samples from shallow boreholes (3 to 25 m deep) in the Canadian Arctic Islands are included in this overall study.

THEORY

An expression often used to calculate the fraction of ice contained in permafrost pore space is that proposed by Timur (1968). This is the three-phase time-average relationship:

$$\frac{1}{V_p} = \frac{\phi(1 - F_i)}{V_1} + \frac{\phi F_i}{V_i} + \frac{(1 - \phi)}{V_m},$$

in which

V_p = P-wave velocity in permafrost;

V_1 = P-wave velocity in pore water;

V_i = P-wave velocity in ice;

V_m = P-wave velocity in solid matrix material;

F_i = fraction of ice in pore space;

ϕ = porosity of permafrost.

This equation, however, applies to consolidated porous rocks and is inapplicable to unconsolidated permafrost sediments, unless an artificially low value of V_m for unconsolidated material is assumed [as suggested by Hoyer et al. (1975)].

The relationship proposed here, that is shown to fit the data obtained for frozen unconsolidated sediments, is based on the Kuster and Toksöz (1974a) two-phase model, modified to account for the three phases found in permafrost: ice, water, and solid matrix material. The effective moduli are given by

$$\frac{K^* - K}{3K^* + 4\mu} = c\,\frac{K' - K}{3K' + 4\mu},$$

$$\frac{\mu^* - \mu}{6\mu^*(K + 2\mu) + \mu(9K + 8\mu)} = \frac{c(\mu' - \mu)}{6\mu'(K + 2\mu) + \mu(9K + 8\mu)},$$

and

$$\rho^* - \rho = c(\rho' - \rho),$$

where the starred are the effective values, the primed the inclusion values, and the unprimed the matrix values; and in which

K = bulk modulus;

μ = shear modulus;

ρ = density;

c = concentration of inclusions.

The first stage in developing the model is to consider an ice matrix in which spherical water inclusions form. The elastic properties are calculated for a complete spectrum of ice/water ratios. The elastic constants for this ice/water mixture are then employed in a second model of a matrix of varying ice/water content with quartz inclusions. Thus the elastic properties for the three-phase system are obtained. Kuster and Toksöz (1974b) have demonstrated the validity of their model for elastic-wave velocities in suspensions of solid particles in liquids, over a wide range of concentrations. Toksöz et al. (1976) have demonstrated its validity for determining the velocities in liquid-saturated porous rocks.

The results of the analysis described are plotted in Figure 1, where compressional-wave velocity, V_p, is plotted as a function of permafrost porosity for a complete range of pore space fractional ice contents. The elastic constants and density employed are for ice: bulk modulus, 8.4 x 10^9 Pa; shear modulus, 3.7 x 10^9 Pa; density, 920 kg/m^3. Those employed for water are: bulk modulus, 2.0 x 10^9 Pa; density, 1000 kg/m^3. Those for quartz are: bulk modulus, 44.0 x 10^9 Pa; shear modulus, 37.0 x 10^9 Pa; density, 2700 kg/m^3.

TEST SPECIMENS AND EXPERIMENTAL PROCEDURES

The thirty-seven permafrost samples were recovered in their natural, frozen state. After sealing in layers of polyethylene film or in plastic tubing, the samples were maintained in their frozen state (at a temperature of approximately -9°C after shipment) during shipment, storage, test-specimen preparation, and until they were tested under controlled-environmental conditions.

Test specimens were prepared in their frozen state following the procedures described by King (1977), except that for these experiments the cylindrical specimens were ground 50 mm in diameter and approximately 50 mm in length. The test procedures employed were as described by King (1977) and Pandit and King (1979), except that the specimens were subjected only to a hydrostatic confining stress of 0.35 MPa (50 Psi) under drained conditions. Earlier experiments with shales of high water content indicated a tendency for specimens to lose interstitial water when axial stresses were superimposed, even at temperatures below 0°C. A hydrostatic stress of 0.35 MPa, under drained conditions, is representative of that experienced by the permafrost samples in their natural state. Kurfurst (1976) and J.A. Hunter (personal communication) have commented on the excellent agreement of the laboratory velocities with those measured in field seismic surveys.

Compressional-wave velocities at frequencies in the range 500 kHz to 850 kHz were measured by a first-pulse arrival technique [described by King (1977)] at ascending temperatures in the range -16° to +5°C, except for those specimens recovered from the Beaufort Sea for which the temperature range was -10° to -1°C [see King et al. (1982) for a full discussion of this series of tests]. Shear-wave velocities were also measured sequentially along with the compressional-wave velocities, but it generally proved difficult to identify shear-wave arrivals unequivocally at the low confining stress employed.

The bulk density, porosity, water content, and grain size distribution were measured by standard geotechnical and soil science procedures (Lambe, 1951) on small samples taken from the samples used to prepare the test specimens. Standard X-ray diffraction analyses were also performed on representative samples to identify the clay minerals present. Kaolinite, illite, and chorite were found in fairly constant proportions of 3:1:1, respectively, for all samples analyzed. The results of estimates of total clay contents from five X-ray analyses are shown in Figures 2 and 3 for the seventeen samples from the Canadian Arctic Islands. The range of total clay contents for the nine representative samples tested was zero to 40 %, but these estimates are subject to a number of possible errors as indicated by Carroll (1970).

EXPERIMENTAL RESULTS

Compressional-wave velocities V_p for 17 permafrost specimens from the Canadian Arctic Islands are shown plotted as a function of ascending temperature in the range -16° to +5°C in Figures 2 and 3. Also indicated in Figures 2 and 3 are the porosity ϕ, particle size fractions, and, for five specimens, the fractional clay content. Figure 2 shows the result for specimens containing less than 40 % clay-sized particles; Figure 3 shows the results for those containing 40 % or more clay-sized particles. Descriptions of the Mackenzie River samples and some V_p measurements have been reported by Kurfurst (1976). Descriptions of the Beaufort Sea samples and V_p measurements have been reported by King et al. (1982).

In all cases the pore spaces of the permafrost specimens were found to be completely occupied by ice and water at the start of the tests. Confirmation was provided by noting bulk density before testing and the reduction in bulk volume and water content of each specimen after removal from the pressure cell at room temperature; typically the reduction in bulk volume was 3-4 %.

The compressional-wave velocities for all 37 permafrost specimens are plotted in Figures 4 and 5 as a function of fractional clay content F_c for temperatures of -10° and -5°C, respectively. Linear regressions for porosities $\phi < 0.40$ and $\phi > 0.40$ are also shown in each figure. In Figure 6, V_p measurements are plotted as a function of porosity (Figure 6a) and identified according to fraction of clay-sized particles (Figure 6b) at a temperature of +5°C.

The fraction F_i of ice contained in the pore space has been calculated for each specimen at temperatures of -10°, -5°, and -2°C from Figure 1, based on the Kuster and Toksöz (1974a) theory discussed earlier. The calculated values of F_i are plotted as a function of fraction F_c of clay-sized particles at temperatures of -10° and -5°C in Figures 7 and 8, respectively. Linear regressions for porosities $\phi < 0.4$ (mean porosity: 0.36; standard deviation: 0.03) and $\phi > 0.4$ (mean porosity: 0.42; standard deviation: 0.02) are also shown in each figure.

V_p measurements for all specimens are plotted in Figure 9 at temperatures of -10°, -5°, and -2°C as a function of water-filled porosity, $\phi(1 - F_i)$. A linear regression of the data is also shown in the figure.

DISCUSSION

It is clear from the V_p measurements plotted in Figures 2 and 3 as a function of temperature that the presence of clay-sized particles in permafrost influences the velocities in two ways: First, an increase in F_c results in a lower measured V_p at a given temperature in the frozen state. Second, values of F_c greater than 0.40 appear to result in a depression of temperature below 0°C at which water in the pore space is unfrozen. The temperature depression for the unconsolidated sediments studied here is approximately 1°C. Similar behavior for shales has been reported previously by Timur (1968) and King (1977).

When the V_p measurements are plotted as a function of fraction of clay-sized particles for a given temperature below 0°C in Figures 4 and 5, the first of the effects referred to above is clearly observed. Linear regressions for porosities $\phi < 0.40$ and $\phi > 0.40$ indicate that an increase in porosity at a given clay content results in a significantly higher measured V_p. It is interesting to note that the V_p measurements for specimens containing low fractions of clay-sized particles are close to V_p values predicted from the Kuster and Toksöz theory plotted in Figure 1 for fractional ice contents F_i close to 1.0.

At temperatures above 0°C it is clear from Figure 6b that there is no correlation of V_p with F_c. There does, however, appear to be a correlation of V_p with ϕ at +5°C, as shown in Figure 6a. V_p calculated for zero ice content from the Kuster and Toksöz (1974a) theory is shown in Figure 6a. It is seen to form a lower bound to the measured velocities. Higher measured V_p values are to be expected for saturated porous sediments, particularly those having low fractions of clay-sized particles, because of the effect on intergranular forces due to the 0.35 MPa confining stress at which the V_p measurements were made. Also shown in Figure 6a is V_p calculated for water-saturated packing of quartz spheres at 0.35 MPa, employing theories [Deresiewicz (1958)] for V_p in regular packings of spheres subjected to confining stress, and Biot's (1962) theory for V_p in liquid-saturated porous media. When ice is present in the pore space at temperatures below 0°C, King and Pandit (1979) have demonstrated that intergranular stresses are considerably less important and the measured velocities are relatively insensitive to changes in confining stress.

The fractional ice contents F_i calculated from the Kuster and Toksoz theory and plotted as a function of fraction F_c of clay-sized particles in Figures 7 and 8 for temperatures of -10° and -5°C indicate that F_i is influenced by the permafrost porosity ϕ. An increase in ϕ for a given F_c results in a significant increase in F_i. Figure 9 shows the compressional-wave velocity V_p plotted as a function of water-filled porosity $\phi(1 - F_i)$. The linear relationship $V_p = 4.14 - 6.23\ \phi(1 - F_i)$ fits the data very well, particularly over the range of values of $\phi(1 - F_i)$ from 0.00 to 0.20, despite the fact that the Kuster and Toksöz (1974a) theory does not predict a unique dependence of V_p on $\phi(1 - F_i)$. There is no statistically significant difference between the data for temperatures of -2°, -5°, or -10°C.

CONCLUSIONS

It is concluded that the compressional-wave velocity for unconsolidated permafrost from the Canadian Arctic correlates well with the water-filled porosity $\phi(1 - F_i)$, irrespective of the original porosity, fraction of clay-sized particles, or temperature, when F_i is calculated from the Kuster and Toksöz theory. This correlation remains to be confirmed by actual measurements of the fractional ice content by an experimental technique such as nuclear magnetic resonance.

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FIGURE CAPTIONS

Figure 1 Compressional-wave velocity for unconsolidated permafrost as a function of porosity, for different ice content.

Figure 2 Measured compressional-wave velocity for permafrost as a function of temperature, for fraction of clay-sized particles (< 2 μm) less than 0.40.

Figure 3 Measured compressional-wave velocity for permafrost as a function of temperature, for fration of clay-sized particles (< 2 μm) greater than 0.40.

Figure 4 Measured compressional-wave velocity for permafrost as a function of fraction of clay-sized particles at a temperature of -10°C.

Figure 5 Measured compressional-wave velocity for permafrost as a function of fraction of clay-sized particles at a temperature of -5°C.

Figure 6 Measured compressional-wave velocity for permafrost as a function of (a) porosity and (b) fraction of clay-sized particles at a temperature of +5°C.

Figure 7 Calculated fraction of ice in pore space as a function of fraction of clay-sized particles for a temperature of -10°C.

Figure 8 Calculated fraction of ice in pore space as a function of fraction of clay-sized particles for a temperature of -5°C.

Figure 9 Measured compressional-wave velocity for permafrost as a function of water-filled porosity.

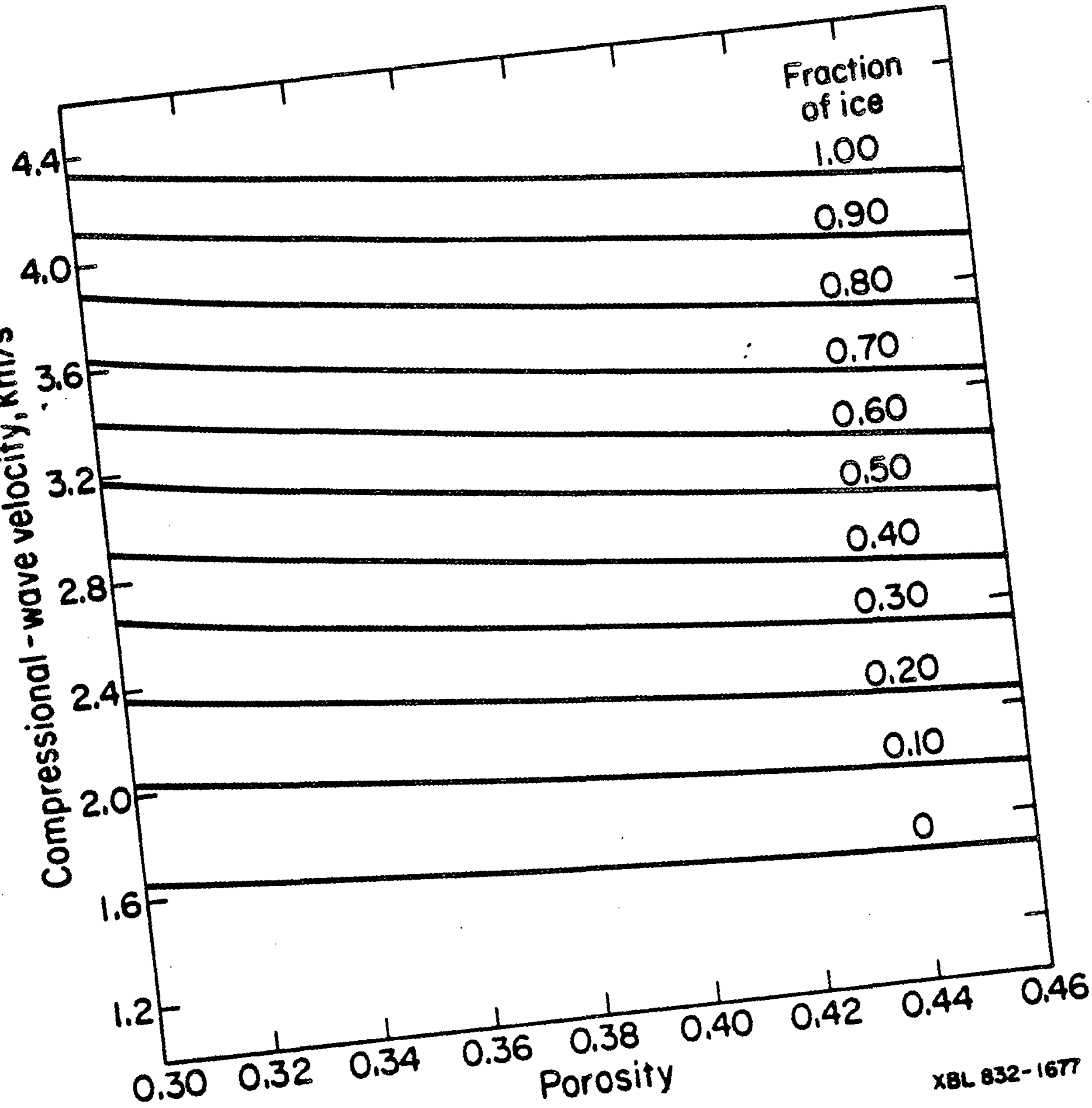

Fraction of ice
1.00
0.90
0.80
0.70
0.60
0.50
0.40
0.30
0.20
0.10
0
Compressional-wave velocity, km/s
4.4
4.0
3.6
3.2
2.8
2.4
2.0
1.6
1.2
0.30
0.32
0.34
0.36
0.38
0.40
0.42
0.44
0.46
Porosity
XBL 832-1677

Figure 1

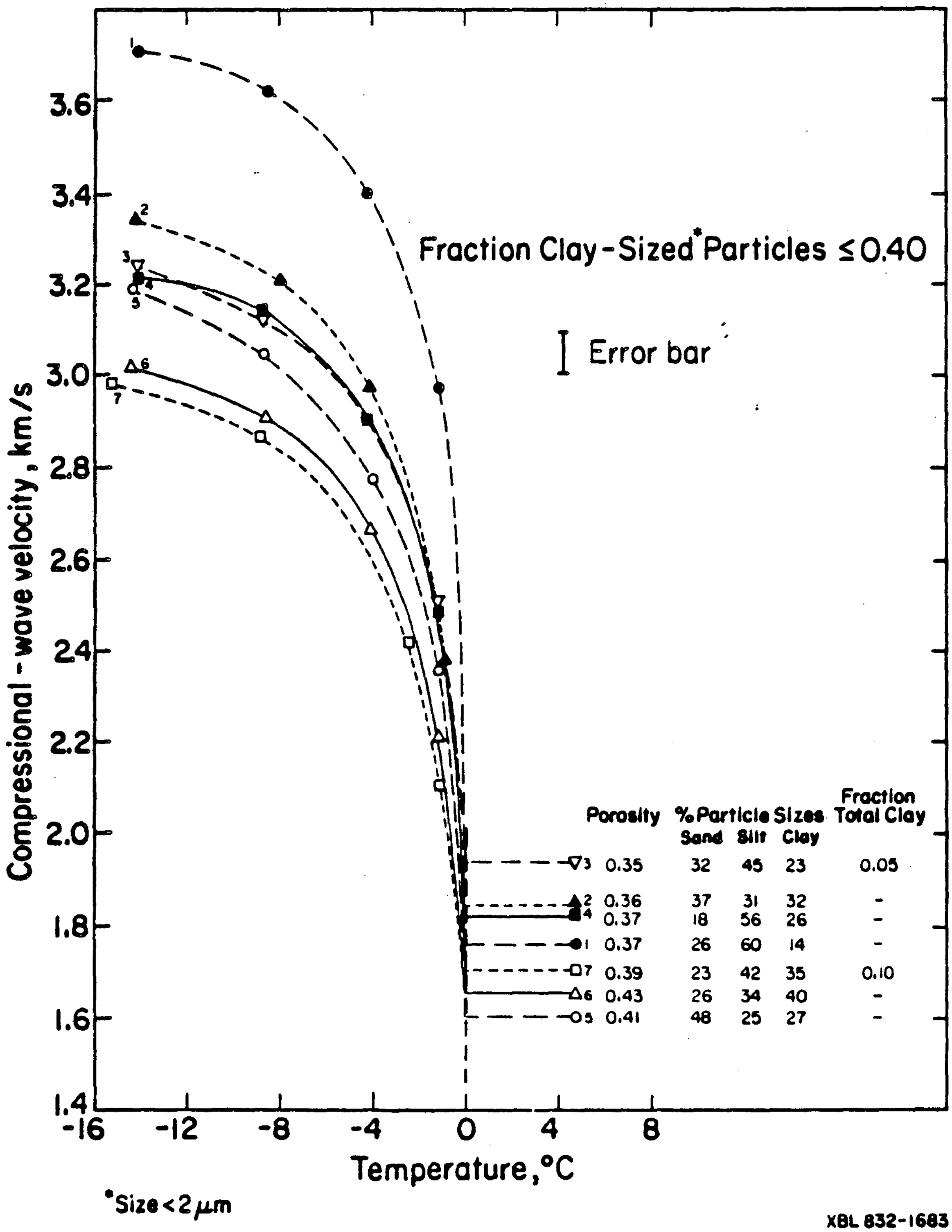
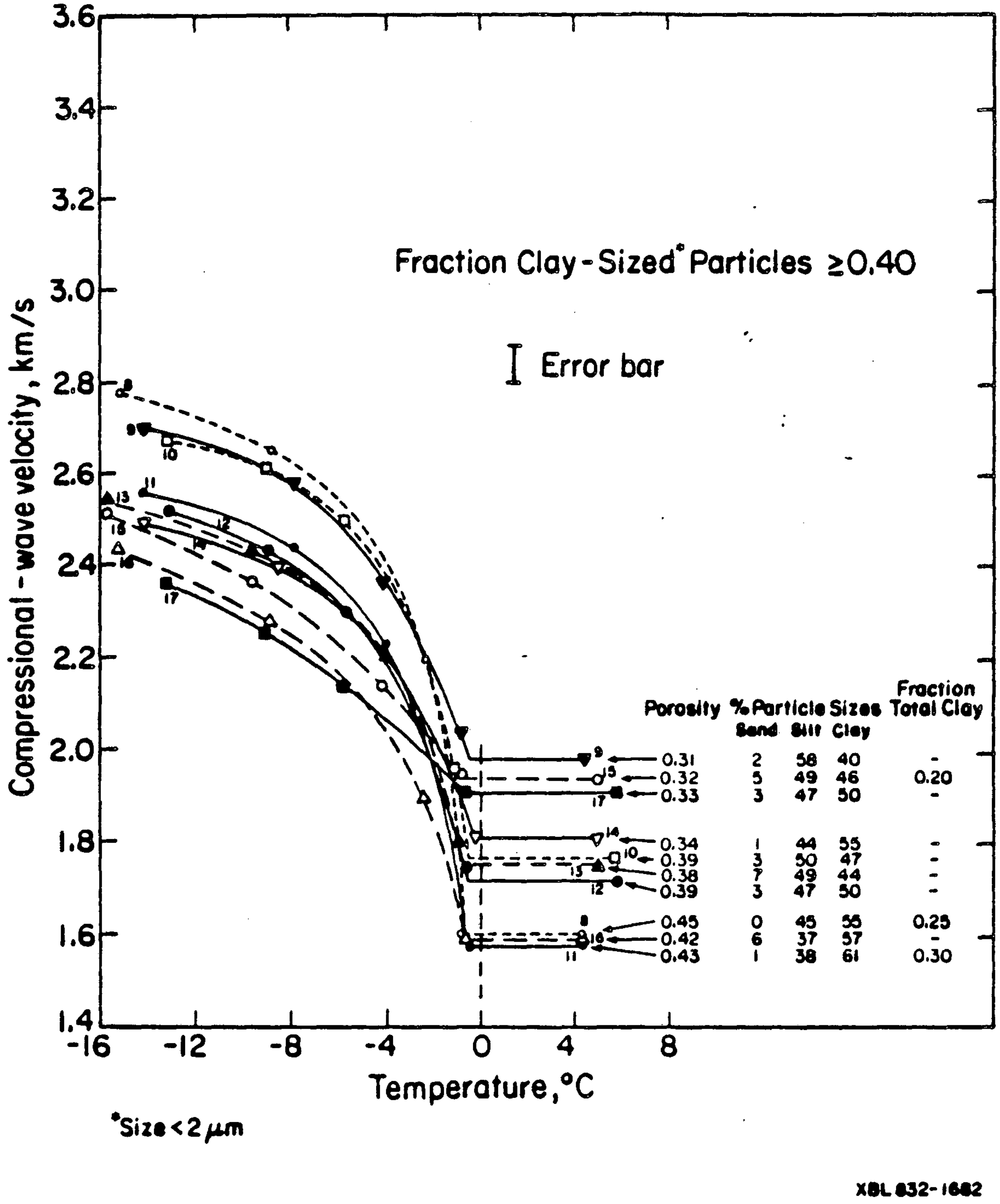
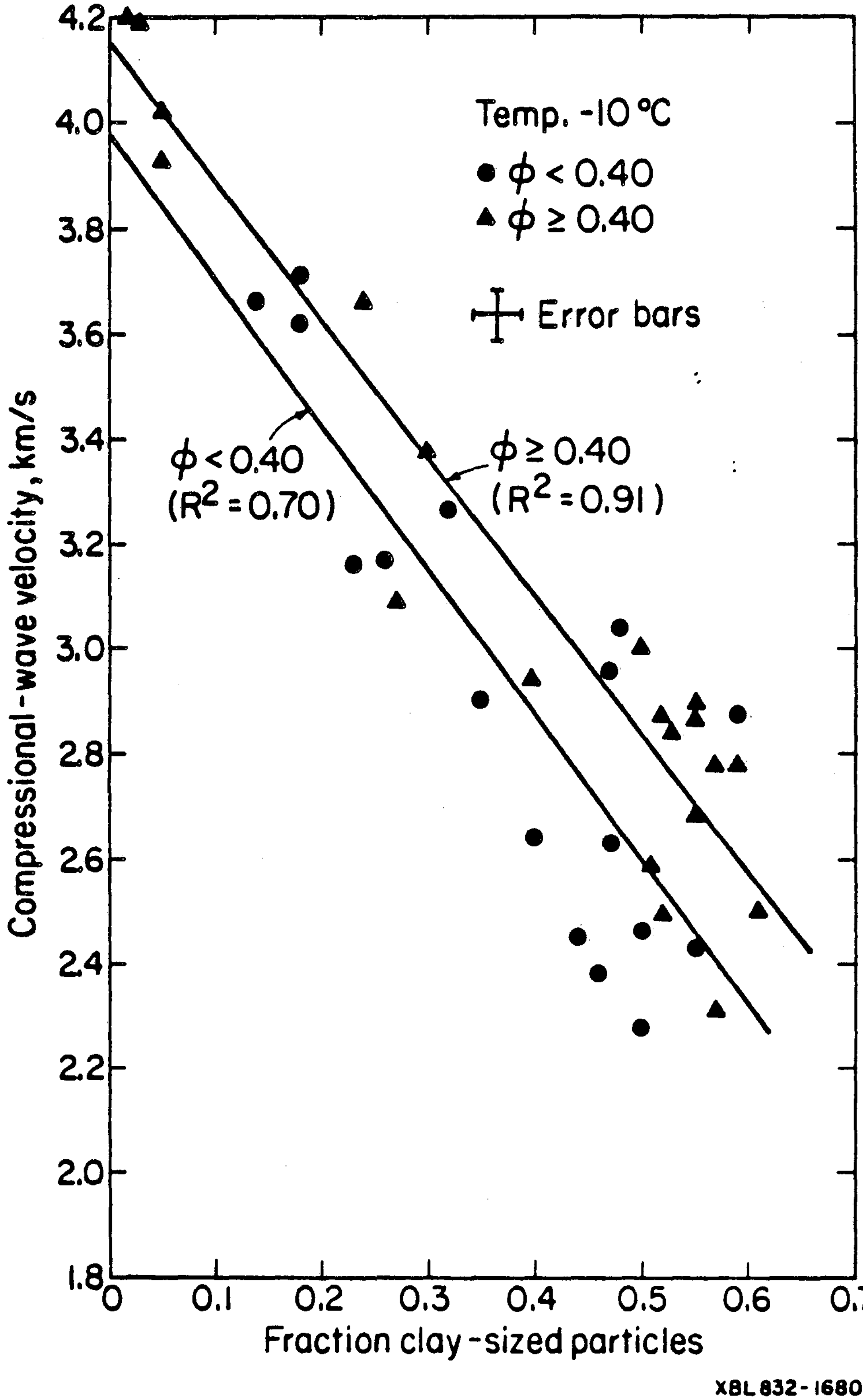

Fraction Clay-Sized* Particles ≤ 0.40
Error bar
Compressional - wave velocity, km/s
Temperature, °C
Porosity
% Particle Sizes
Sand Silt Clay
Fraction Total Clay
3 0.35 32 45 23 0.05
2 0.36 37 31 32 -
4 0.37 18 56 26 -
1 0.37 26 60 14 -
7 0.39 23 42 35 0.10
6 0.43 26 34 40 -
5 0.41 48 25 27 -
*Size < 2 μm
XBL 832-1683

Figure 2



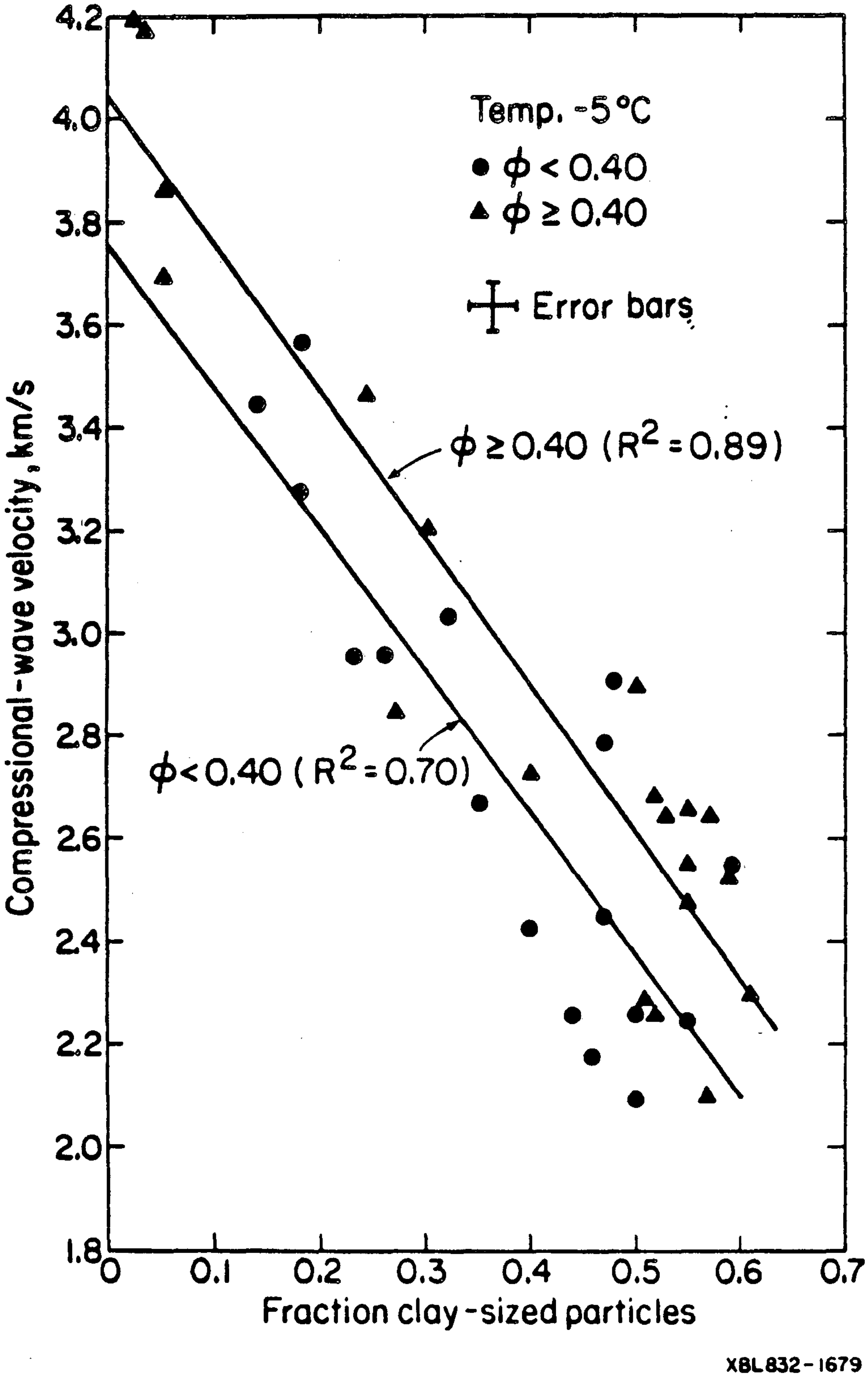
Fraction Clay-Sized* Particles ≥0.40
Error bar
Compressional-wave velocity, km/s
Temperature, °C
Porosity
% Particle Sizes
Sand Silt Clay
Fraction Total Clay
0.31 2 58 40 -
0.32 5 49 46 0.20
0.33 3 47 50 -
0.34 1 44 55 -
0.39 3 50 47 -
0.38 7 49 44 -
0.39 3 47 50 -
0.45 0 45 55 0.25
0.42 6 37 57 -
0.43 1 38 61 0.30
*Size < 2 μm
XBL 832-1682

Figure 3



Temp. -10 °C
● φ < 0.40
▲ φ ≥ 0.40
Error bars
φ < 0.40
(R² = 0.70)
φ ≥ 0.40
(R² = 0.91)
Compressional-wave velocity, km/s
4.2
4.0
3.8
3.6
3.4
3.2
3.0
2.8
2.6
2.4
2.2
2.0
1.8
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
Fraction clay-sized particles
XBL 832-1680

Figure 4



4.2
4.0
3.8
3.6
3.4
3.2
3.0
2.8
2.6
2.4
2.2
2.0
1.8
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
Compressional-wave velocity, km/s
Fraction clay-sized particles
Temp. -5°C
φ < 0.40
φ ≥ 0.40
Error bars
φ ≥ 0.40 (R² = 0.89)
φ < 0.40 (R² = 0.70)
XBL832-1679

Figure 5

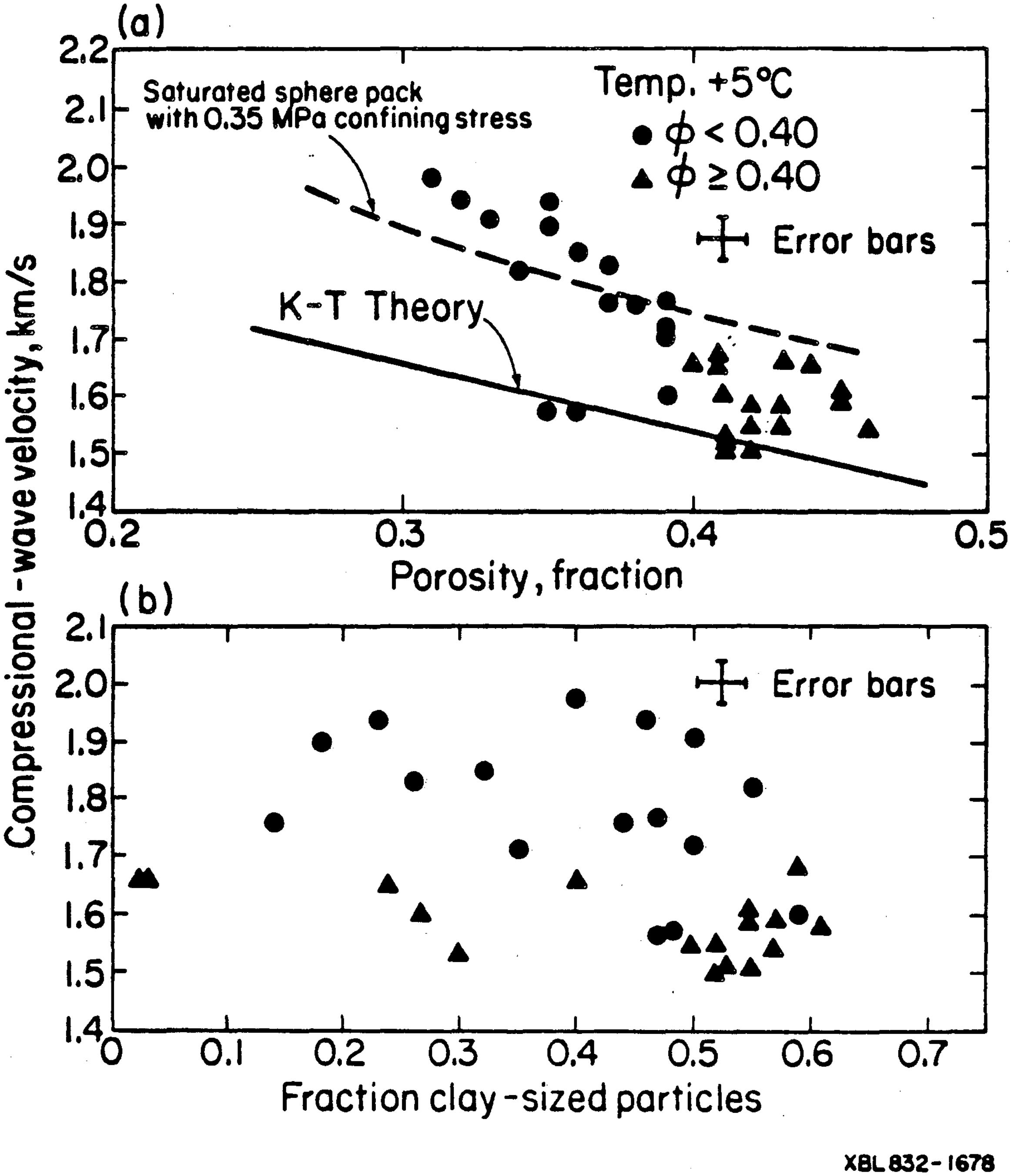

(a)
Saturated sphere pack
with 0.35 MPa confining stress
Temp. +5°C
φ < 0.40
φ ≥ 0.40
Error bars
K-T Theory
2.2
2.1
2.0
1.9
1.8
1.7
1.6
1.5
1.4
0.2
0.3
0.4
0.5
Porosity, fraction
(b)
Error bars
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
Fraction clay-sized particles
Compressional-wave velocity, km/s
XBL 832-1678

Figure 6

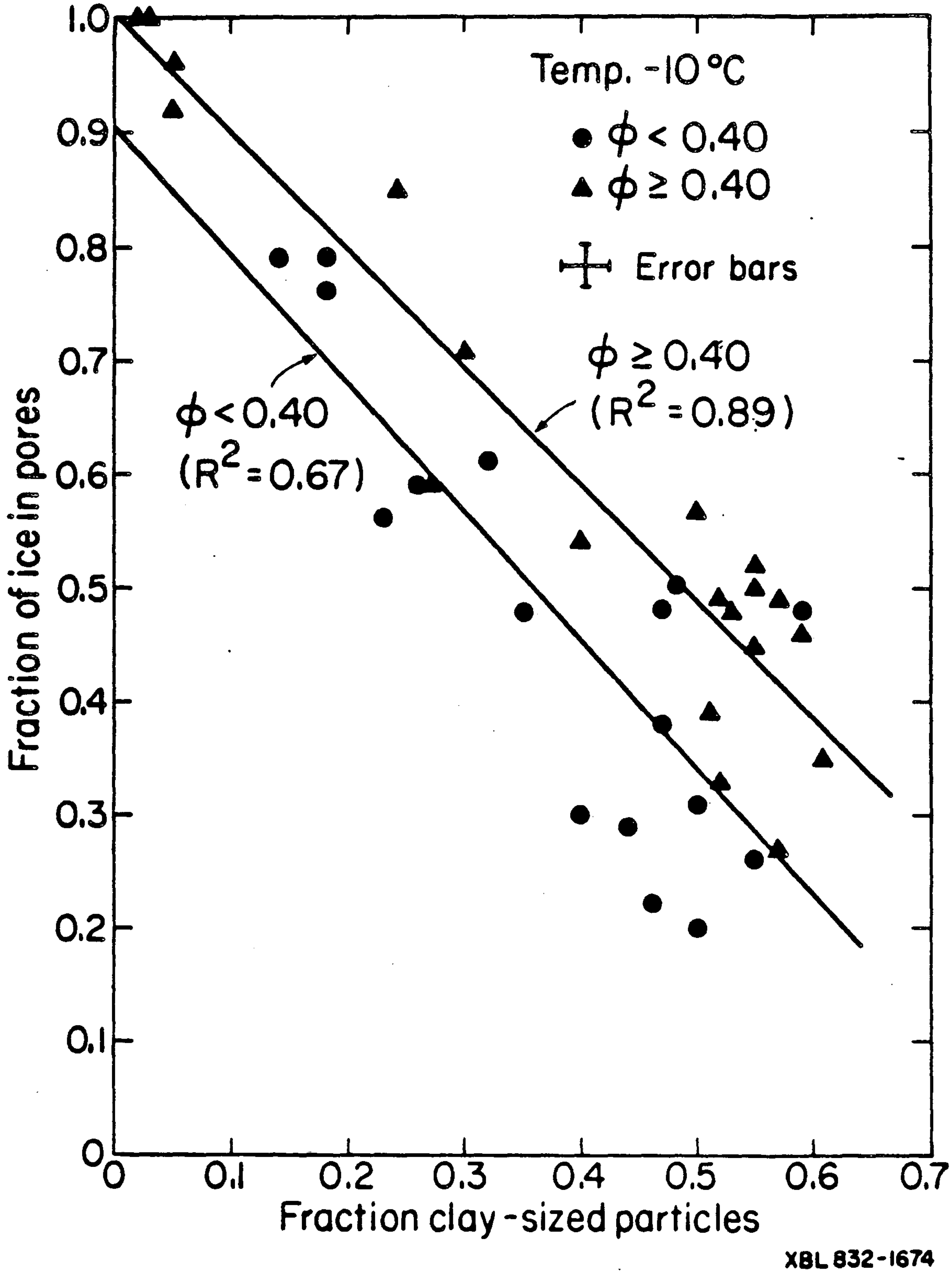
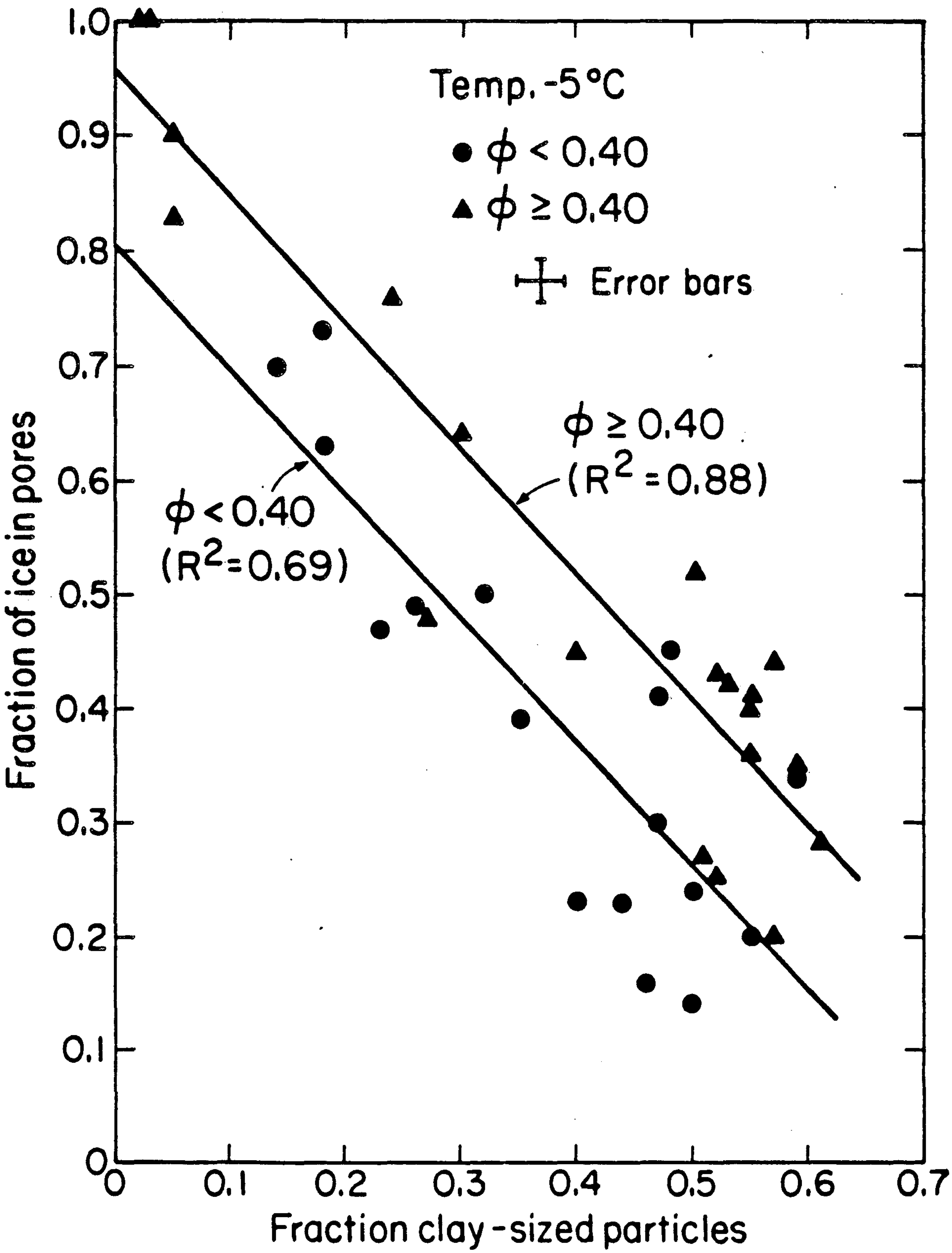
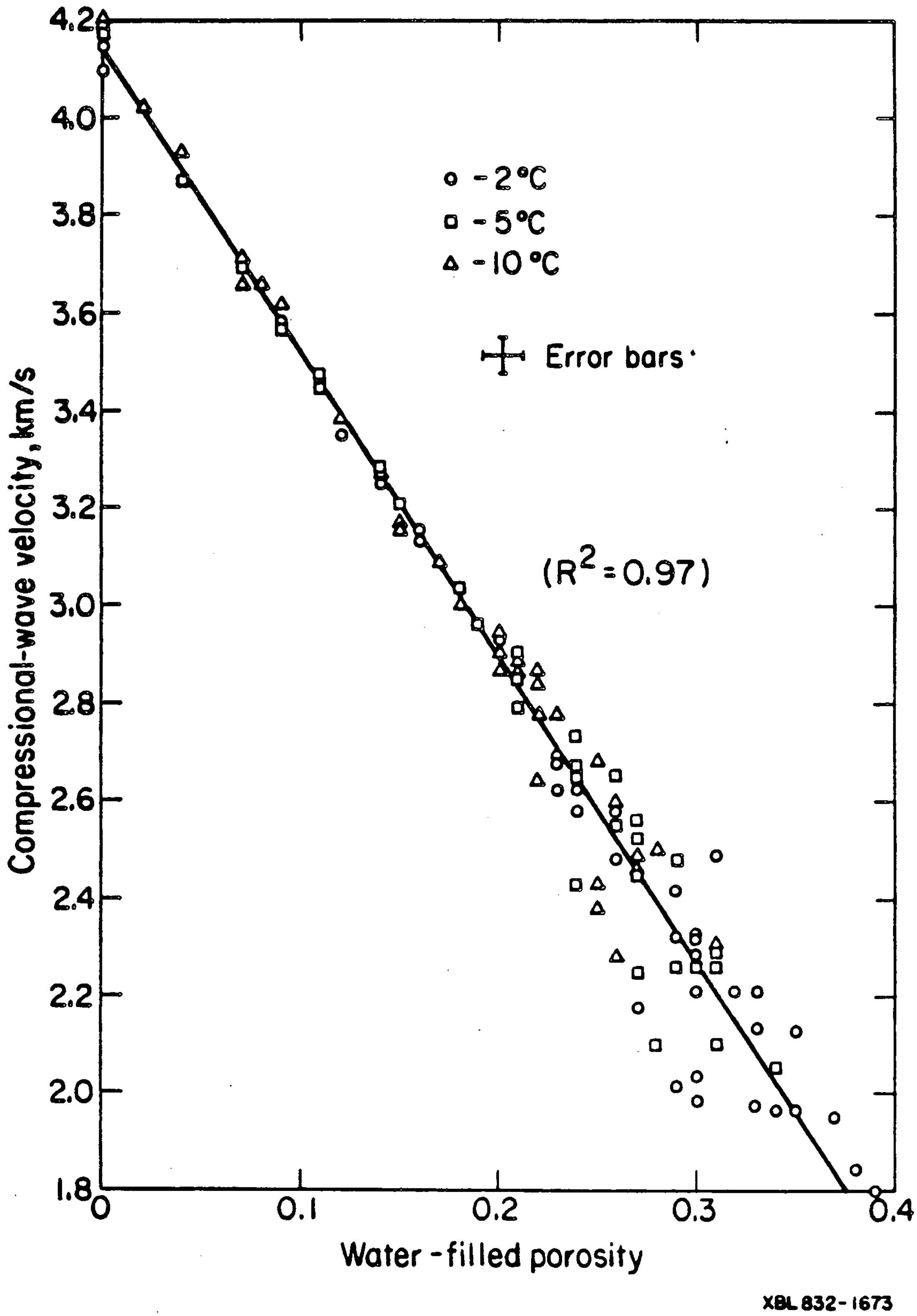

Temp. -10 °C
● φ < 0.40
▲ φ ≥ 0.40
Error bars
φ ≥ 0.40
(R² = 0.89)
φ < 0.40
(R² = 0.67)
Fraction of ice in pores
Fraction clay-sized particles
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
XBL 832-1674

Figure 7



XBL 832-1675

Figure 8



4.2
4.0
3.8
3.6
3.4
3.2
3.0
2.8
2.6
2.4
2.2
2.0
1.8
Compressional-wave velocity, km/s
○ -2°C
□ -5°C
△ -10°C
Error bars
(R^2 = 0.97)
0
0.1
0.2
0.3
0.4
Water-filled porosity
XBL 832-1673

Figure 9

TECHNICAL INFORMATION DEPARTMENT
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UNIVERSITY OF CALIFORNIA
BERKELEY, CALIFORNIA 94720