ELECTRICAL CONDUCTIVITY OF ROCK SAMPLES SUBJECTED TO HIGH TEMPERATURES AND PRESSURES

VOLUME 2

RESULTS, DISCUSSION AND CONCLUSIONS

by

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Given at the 12th AIRAPT Conference, 1989,
Submitted to the Journal of High Pressure Physics.

PAPER 3: GLOVER, P.W.J. et al., 1989 383
Measurements of Saturated Basic and Acidic Rock Electrical Conductivities at Lower Crustal Temperatures and High Pressures.
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CHAPTER 8 : RESULTS

8.1. INTRODUCTION

This chapter contains the graphs related to the confining pressure, pore pressure and temperature variations of rock conductivity of the samples shown in table T8.1. Firstly, however, there is a brief note on the conversion of measured conductance values to rock conductivities and concerning transient drops. These sections are followed by the data.

TABLE T8.1. AN OVERVIEW OF THE SAMPLES STUDIED

IVD1	AMPHIBOLITE
IVD2	"
IVE1	BASIC GRANULITE
IVE2	"
MER21	ORTHOPYROXENITE
MER22	"
L32	AMPHIBOLITE GNEISS
L33B	GRANODIORITE
V1	GRANITE
A8B	AMPHIBOLITE GNEISS
GGRAN1	GRAPHITE GRANULITE
GGRAN2	"
GGRAN3	"
GGRAN4	"

8.2. CONVERSION OF CONDUCTANCE MEASUREMENTS TO CONDUCTIVITY VALUES

It was recognised early in the study that the use of a complex electrode arrangement might make the accurate calculation of the geometric factor of the measurement arrangement difficult. It is this geometric factor that is required to convert measured conductance values into rock conductivities. It has already been shown in chapter 6 that quite an accurate value for the geometric

factor of the new experimental cell has been obtained from comparison measurements of the same rock measured with the new cell and also measured with the Lee Cell. The simplicity of the Lee arrangement has made it possible to obtain a direct comparison of conductivity with the conductance obtained with the new cell and hence enable a calculation of the geometric factor of the new cell to be made. There are, however, several points concerning the experiments to calculate the geometric factor that need to be borne in mind;

A. The comparison was done with samples that had a large porosity (pyrophyllite) and small porosity (amphibolite gneiss). This was so that the effect of porosity upon the geometric factor could be studied.

B. Two samples were required from each of the rocks. This was so that one could be used to characterise the Lee Cell then be taken and inserted into the new cell and the other could be used in the new cell first followed by the Lee Cell. By comparison of these two it was hoped to negate any systematic errors arising from making comparisons between serially processed samples.

C. The comparisons of the two cells were made up to 0.4 GPa and 523 K. Subsequent use of the new cell has been in the range of pressure up to 0.2 GPa and temperatures up to 1173 K. It has had to be assumed that the geometric factor does not change significantly

as the temperature is extended from 523 K to 1173 K as direct comparisons cannot be made in this range due to the limits of the Lee Cell. Extrapolation of the the geometric factor up to 1173 K showed no significant change.

D. The geometry (diameter and length) of the sample was kept unchanged throughout the geometric factor determining experiments and all subsequent experiments.

8.3. TIME VARIATIONS

The temporal variation of conductivity has been noted by many previous authors (Lee, 1977; Evans, 1980; Jackson, 1984) and was noted in all the samples studied. The temporal variation induces what was called a 'hysteresis effect' by Lee (1977). The effect has been discussed in detail in chapter 2. The decrease in the measured conductivity as a function of time is especially marked at high confining pressures and temperatures. The behaviour of initial sample measurements tending to a conductance value but not reaching a completely stable value within a practicable time requires the use of a measurement criterion by which the time for taking the measurements can be judged. Lee defined his measurement criterion in the following way; a measurement was taken when "two successive conductance values fell within 2.5% of each other in a 10 minute period". This criterion has subsequently been used by other authors (Evans, 1980). Evans

showed that such a criterion is likely to produce conductance measurements as much as 40% greater than the values obtained subsequent to holding the samples for a period at high confining pressures and temperatures. Comparisons were done between pre-transient drop measurements and post-transient drop measurements in this work. This will be discussed at greater length in chapter 9 but it is sufficient here to mention that the measurements using the Lee criterion were as much as an order of magnitude too large. The measurements made by Lee took between 25 and 60 minutes to obey his criterion. The criterion defined measurements fall in the range of possible values obtainable between those from the immediate measurement of a rock that has not had time to respond to a particular change in pressure and temperature environment and the post-transient state measurement where all cracks closeable for a particular condition array (P_c, P_p, T) have closed. It is the fully developed post-transient state that is most likely to exist for rocks constrained at depth within the crust and is therefore the most important state in which the samples should be measured if the measurements are truly to simulate lower crustal conditions. Stable values of the conductance of the post-transient state can be obtained by leaving a sample at high temperature and pressure until they reach a stable crack closure state. This was demonstrated by measuring values of conductance variation as a function of confining pressure both before and after high temperature and pressure had been applied to it for some considerable time. Further details are contained in chapter 9.

A criterion is also required when varying the pore pressure. It was found that time was required to enable the saturating fluid to percolate through and open up sutured pores and cracks. This process seems to take longer than the process of closing of the cracks.

The criteria adopted in this work for making pre-transient drop measurements is the same as that adopted by Lee. No criterion was needed for post-transient measurements as these were temporally stable.

8.4. RESULTS

8.4.1. NOTES CONCERNING THE EXHIBITION OF THE RESULTS

The results are divided into four sections:

- Variation of conductivity with confining pressure
- Variation of conductivity with pore pressure
- Variation of conductivity with temperature
- Variation of conductivity as three dimensional and parametric representations of pressure and temperature.

These sections are subdivided by sample number. In the case of temperature variations each sample is represented by three $\log_{10}$(conductivity) graphs. One for the initial temperature variation, one for the repeatable temperature variation and the third for the repeatable temperature variation plotted against 1000/T(K).

Figures F8.1. - F8.14	Confining pressure variation
Figures F8.15 - F8.26	Pore pressure variation
Figures F8.27 - F8.62	Temperature variation
Figures F8.63 - F8.77	Three dimensional and parametric graphs

Annotation of graphs has been kept as simple as possible. The main title gives the name of the sample as given in table T8.1. followed by a code;

PC	Confining pressure variation
PP	Pore pressure variation
T	Repeatable temperature variation
TT	Initial temperature variation
TK	Conductivity as a function of 1000/T(K)
PT	Confining pressure variation at raised temperatures (shown parametrically)
3D	variation as a function of temperature and pressure shown in three dimensions

8.4.2. THE VARIATION OF ROCK ELECTRICAL CONDUCTIVITY AS A FUNCTION OF CONFINING PRESSURE

FIG. F8.1. IVD1PC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

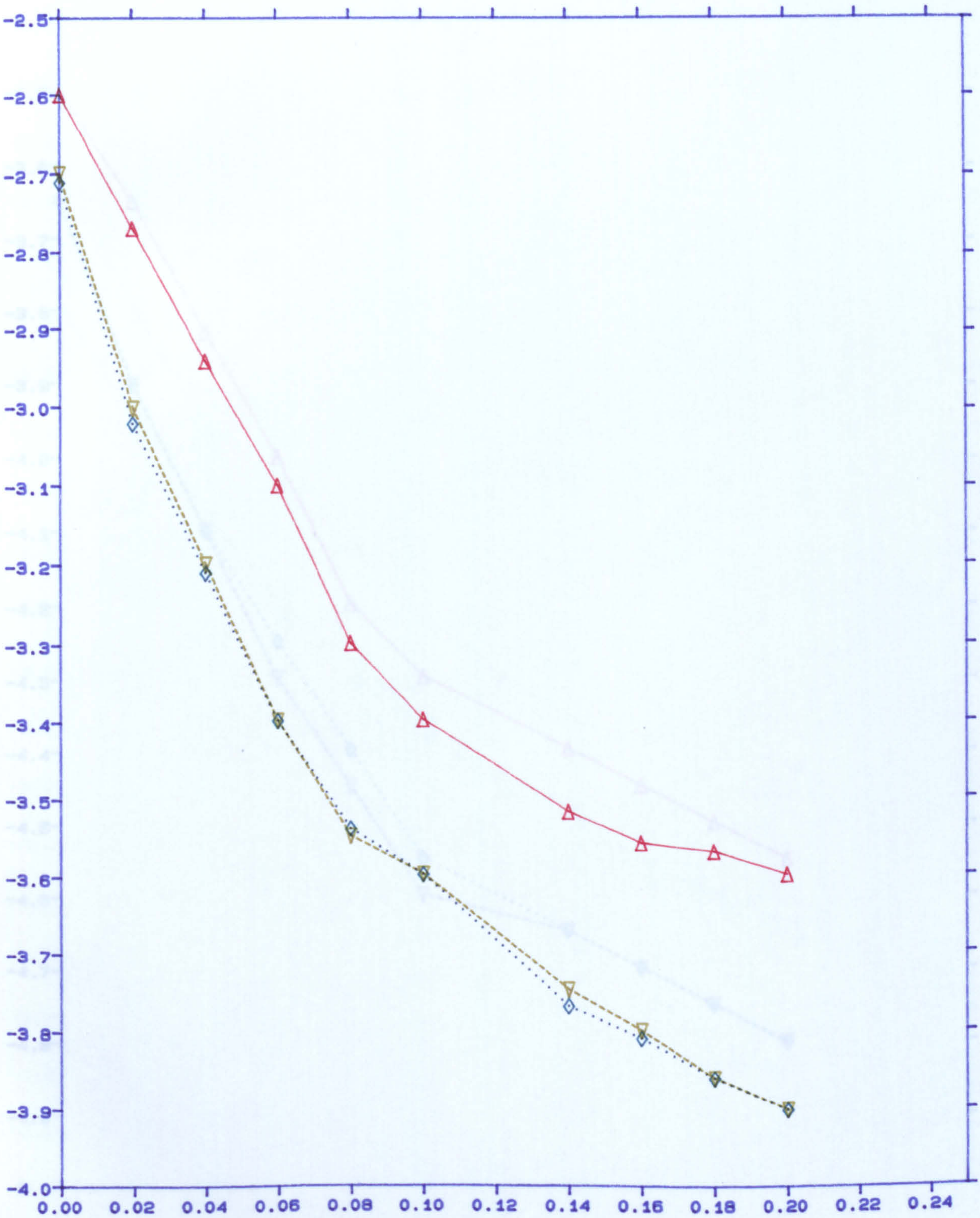


FIG. F8.2. IVD2PC CONFINING PRESSURE VARIATION FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

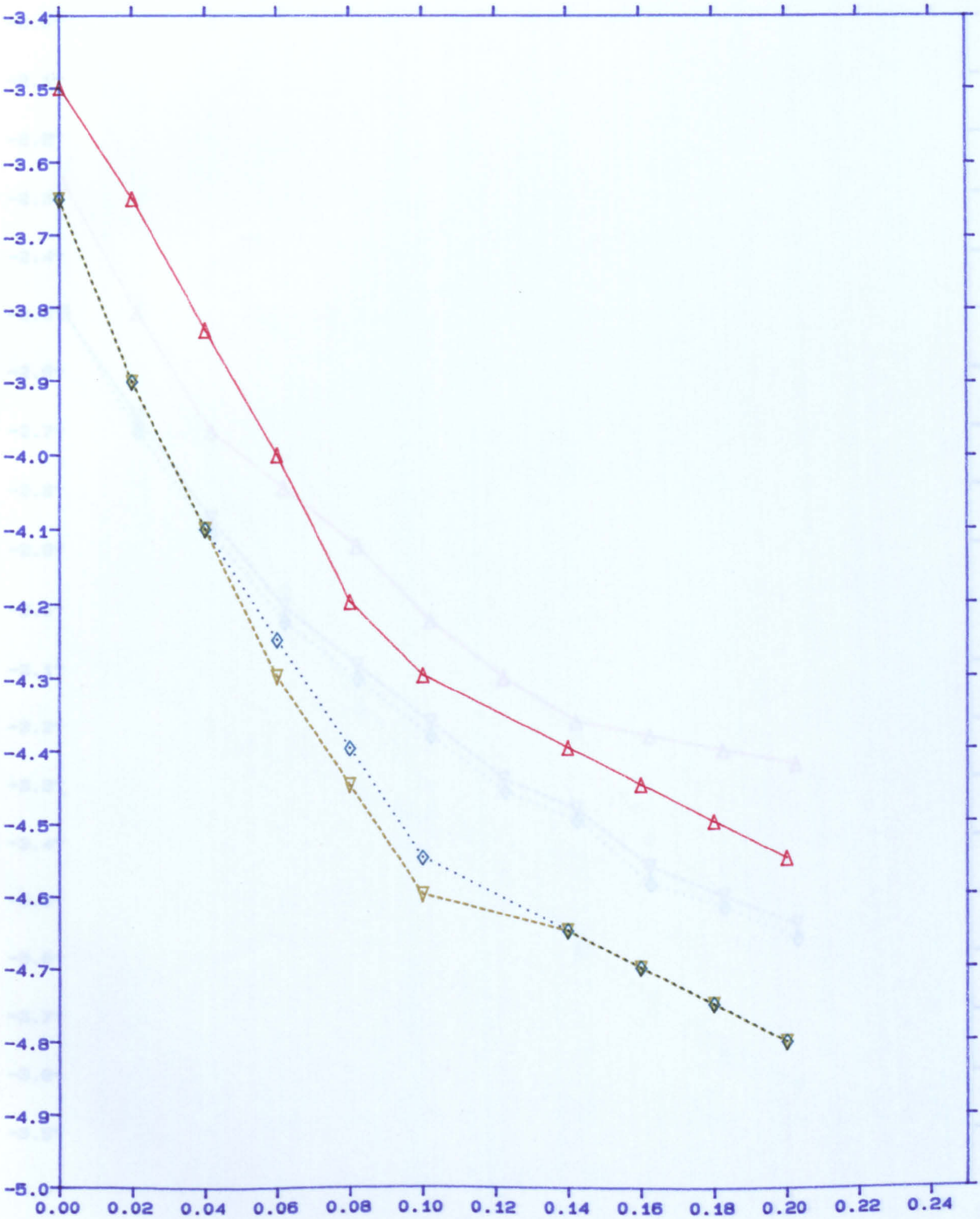


FIG. F8.3. IVE1PC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

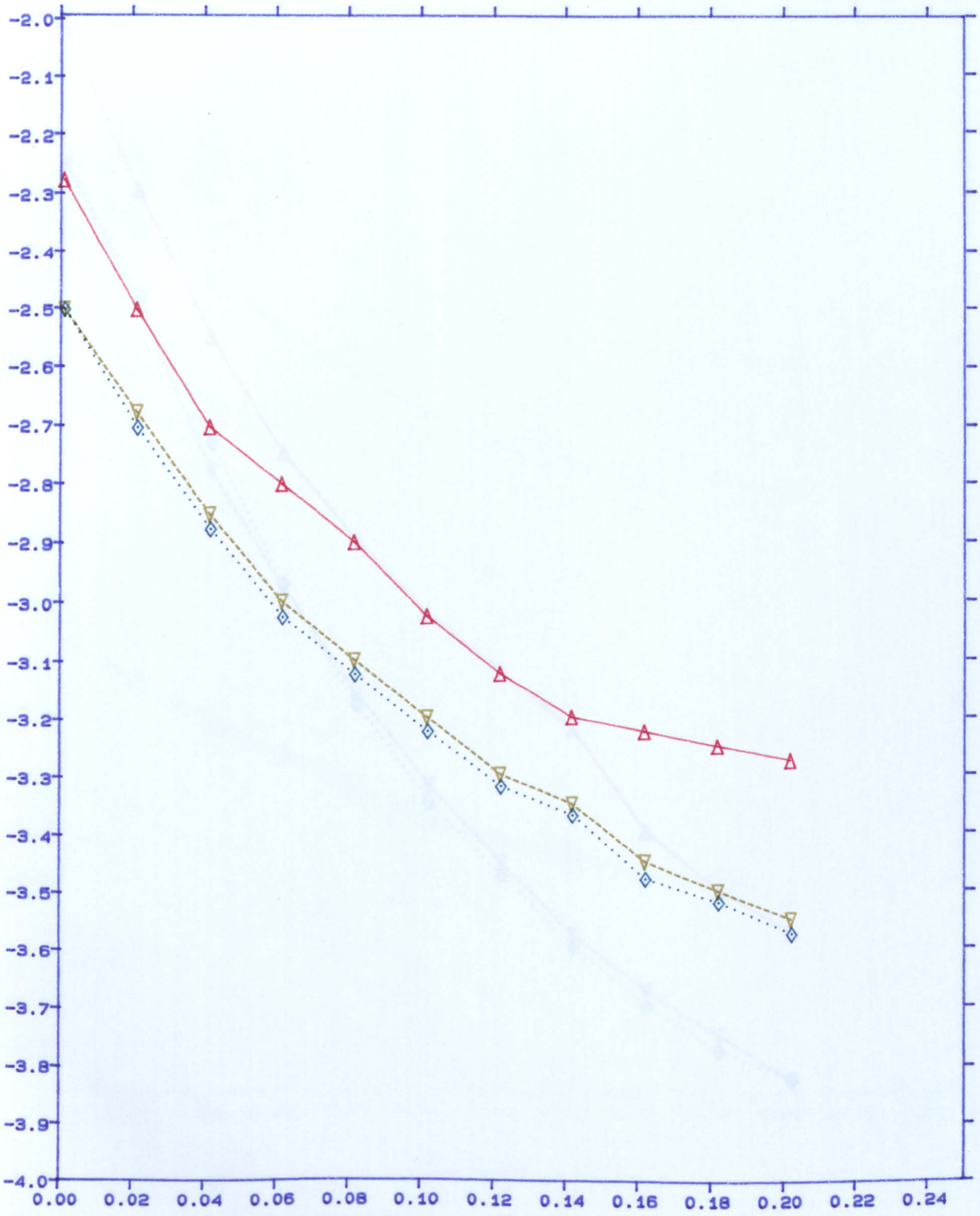


FIG. F8.4. IVE2PC CONFINING PRESSURE VARIATION FOR DISTILLED WATER SATURATED BASIC GRANULITE

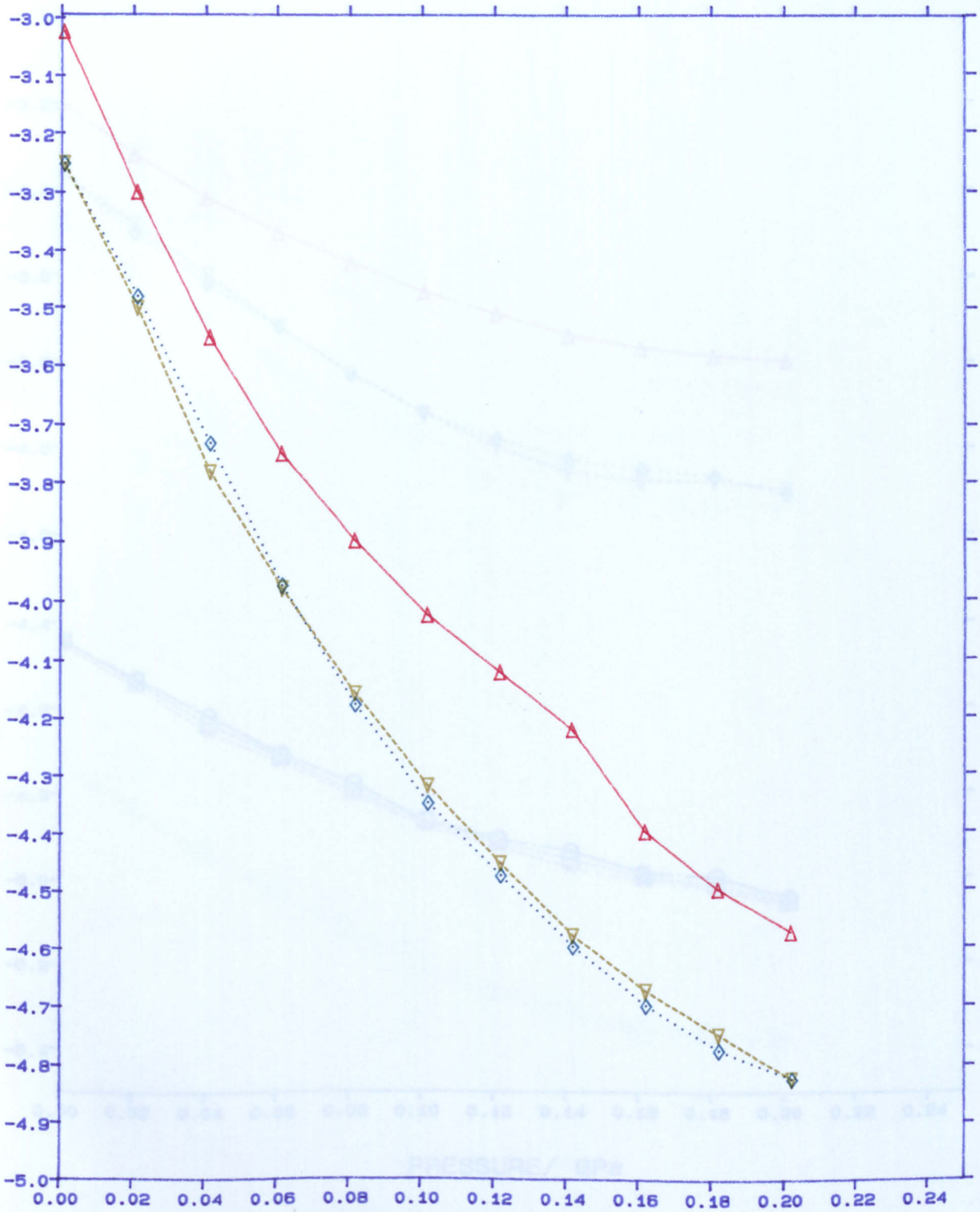
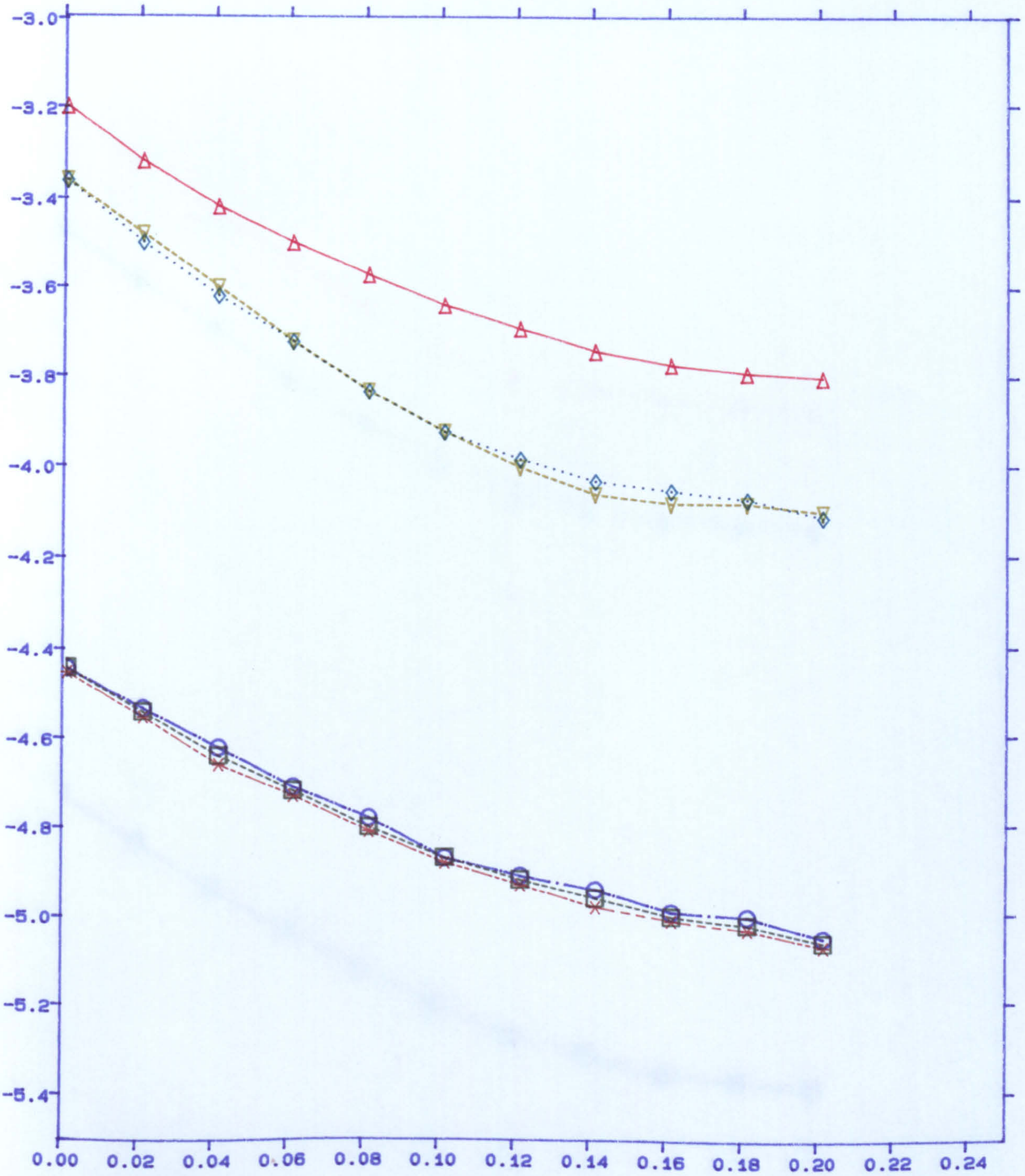
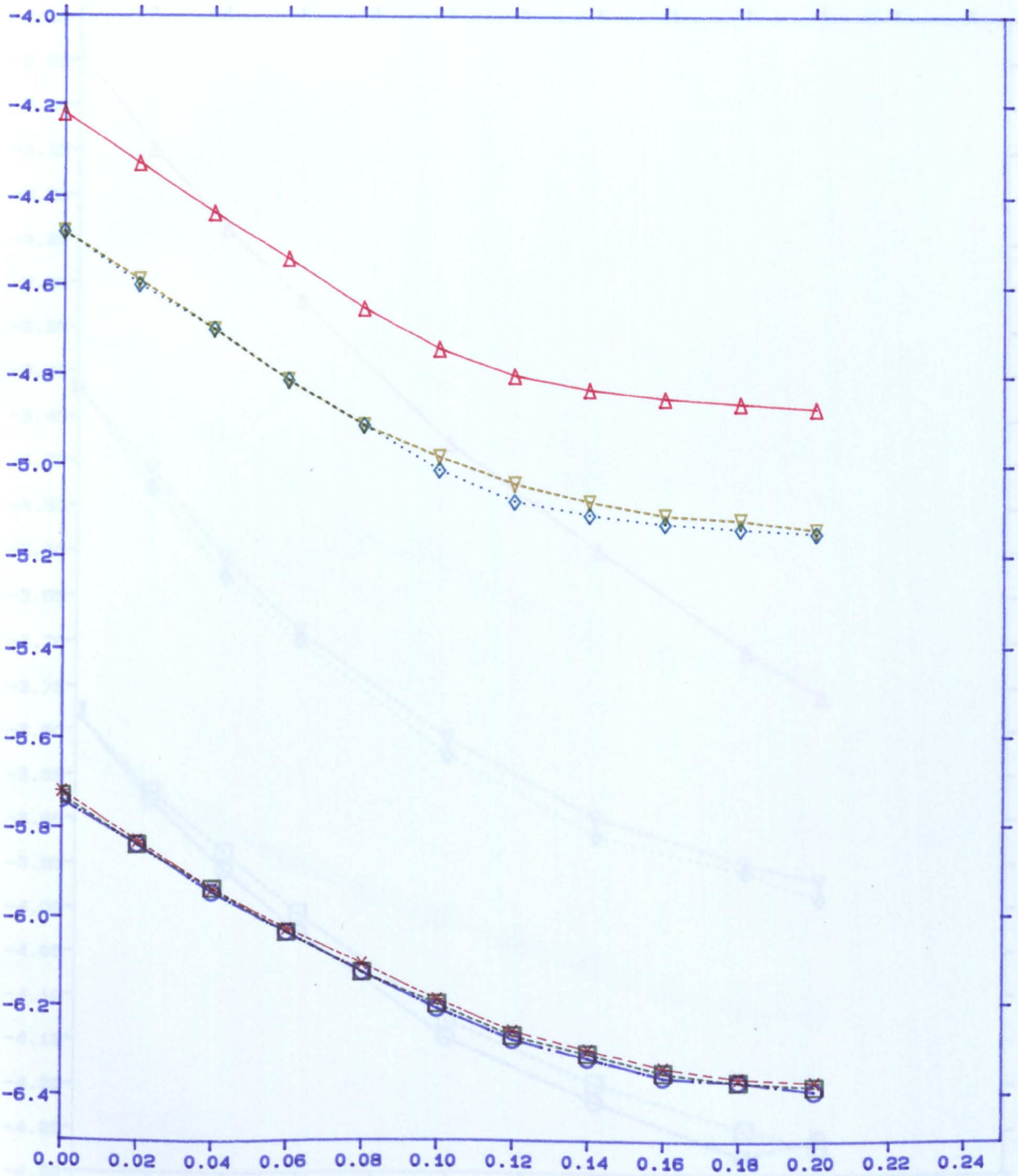


FIG. F8.5. MER21PC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE



INITIAL PRESSURE VARIATION
POST-PRESSURE TRANSIENT VARIATION
POST-PRESSURE TRANSIENT VARIATION
POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F8.6. MER22PC CONFINING PRESSURE VARIATION FOR DISTILLED WATER SATURATED ORTHOPYROXENITE



—Δ— INITIAL PRESSURE VARIATION
----▽---- POST-PRESSURE TRANSIENT VARIATION
·····◇····· POST-PRESSURE TRANSIENT VARIATION
----□---- POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
—○— POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
----×---- POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F8.7. A8BPC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

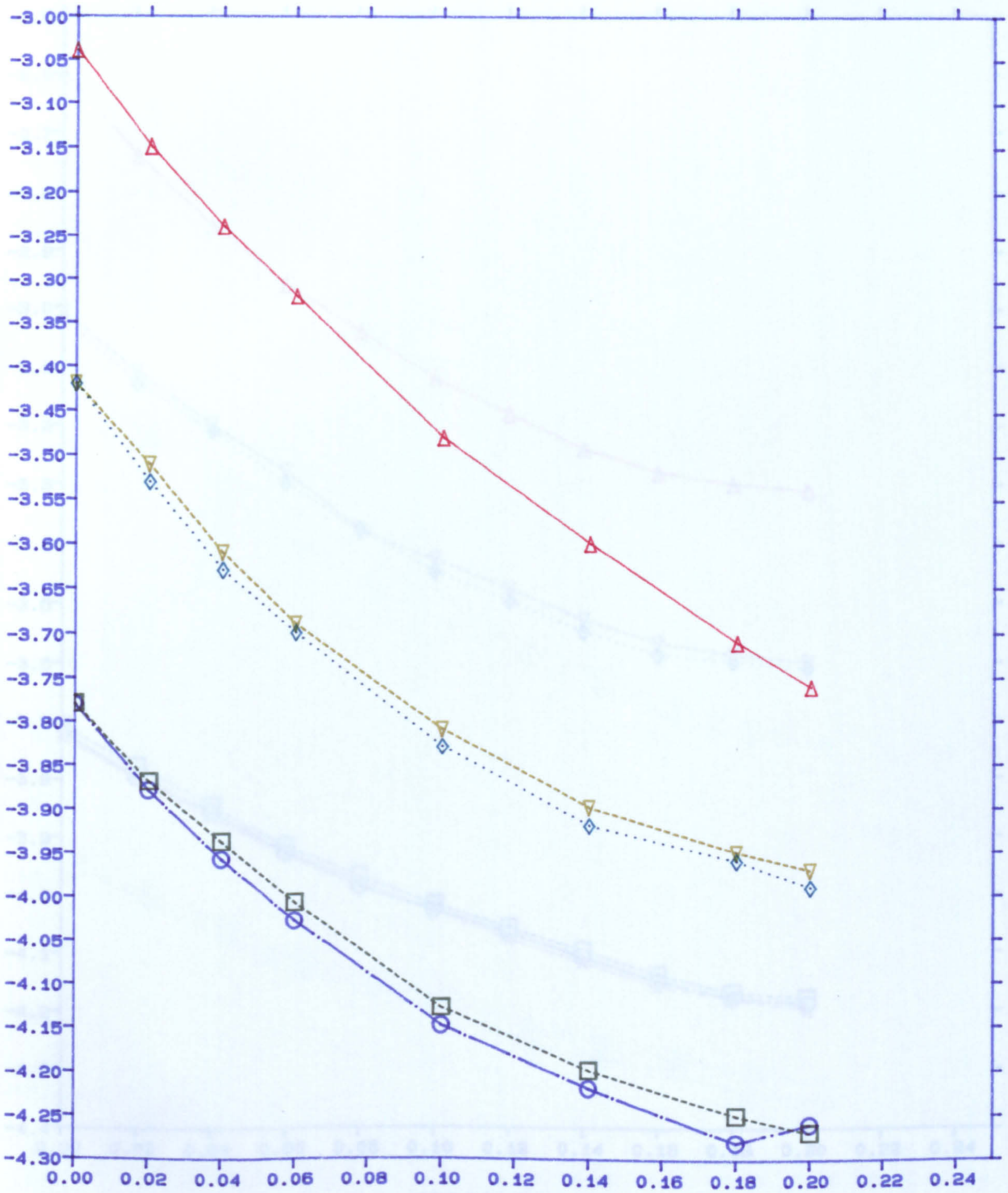


FIG. F8.8. L33BPC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE

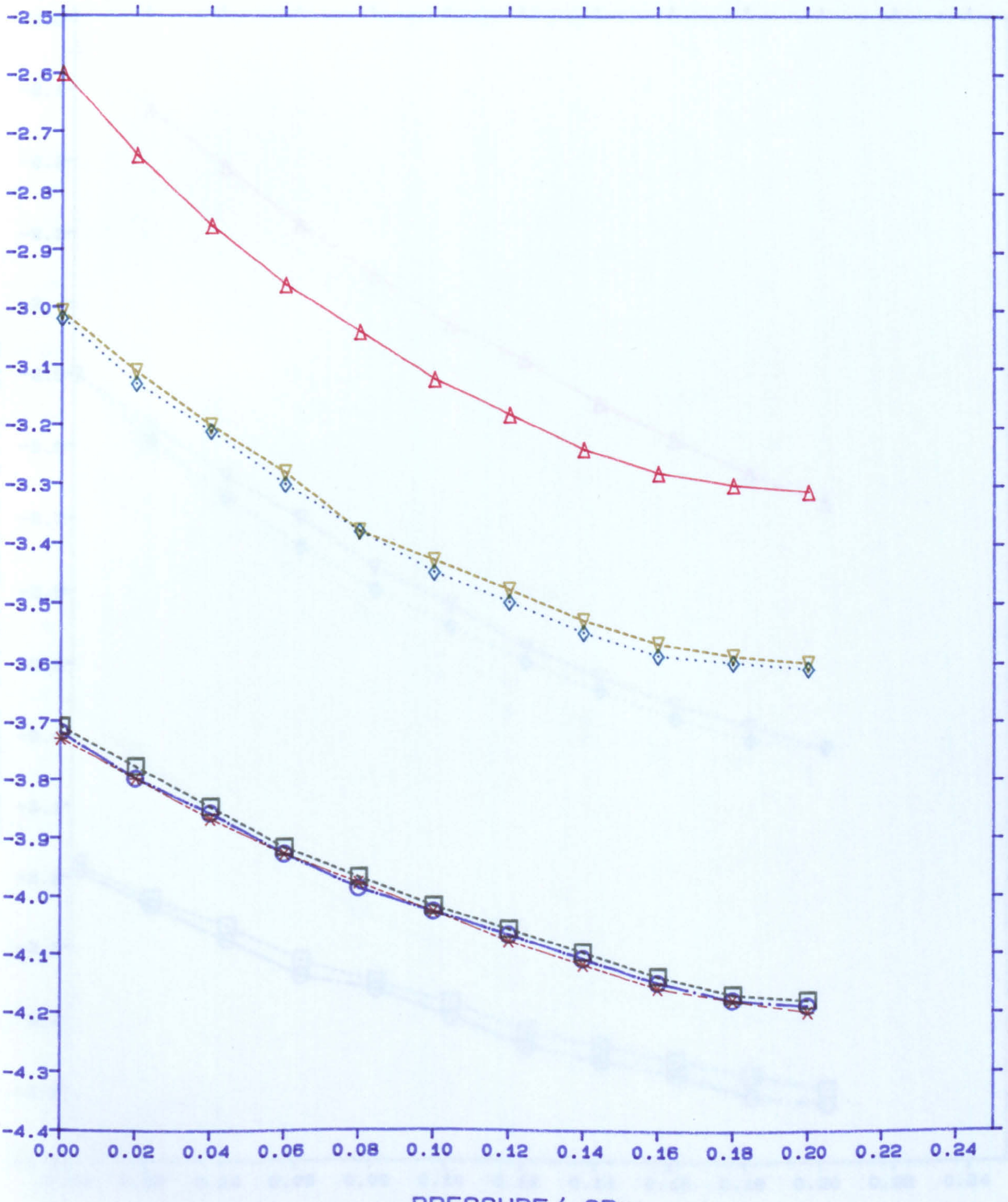


FIG. F8.9. L32PC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

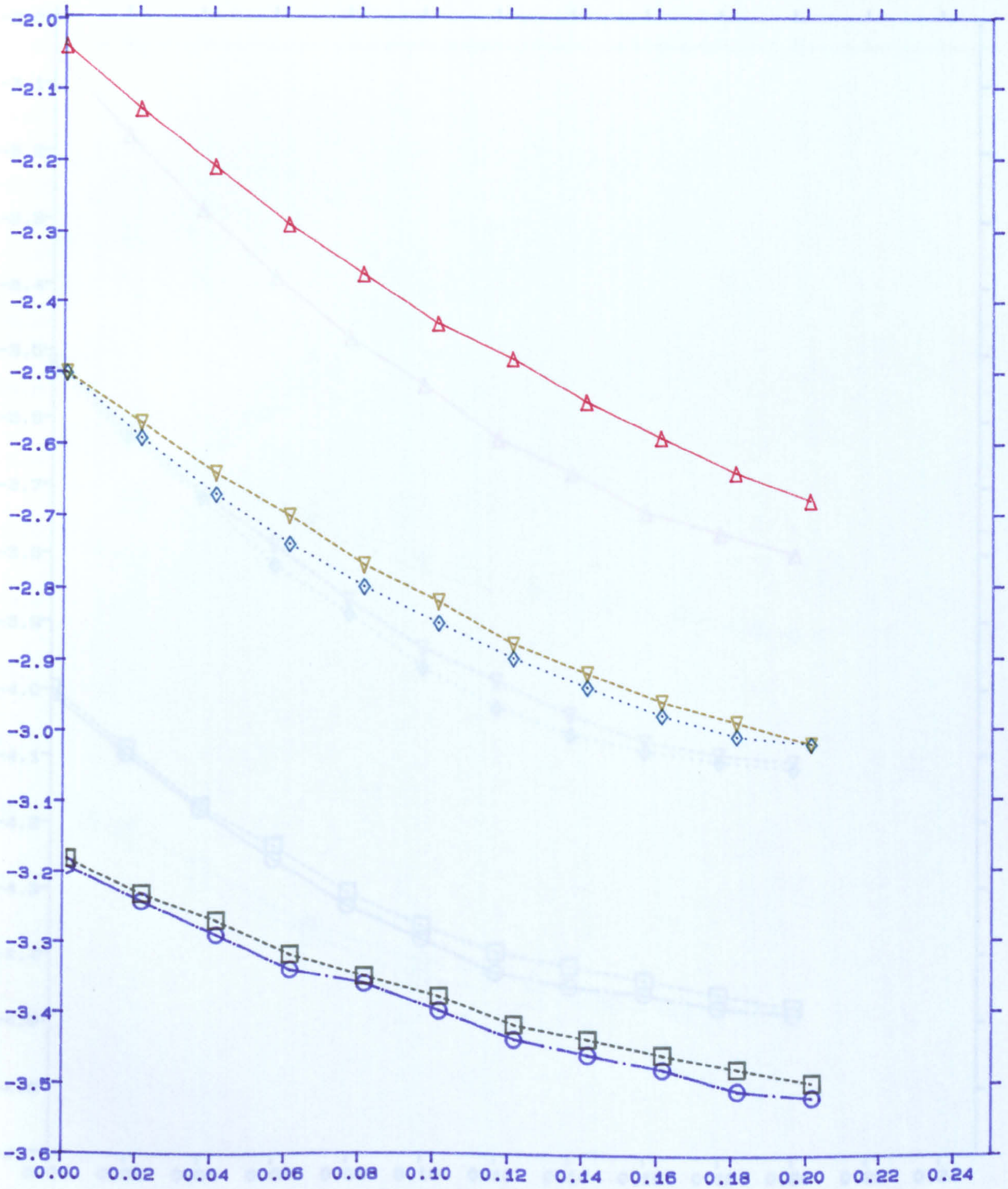


FIG. F8.10. V1PC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANITE

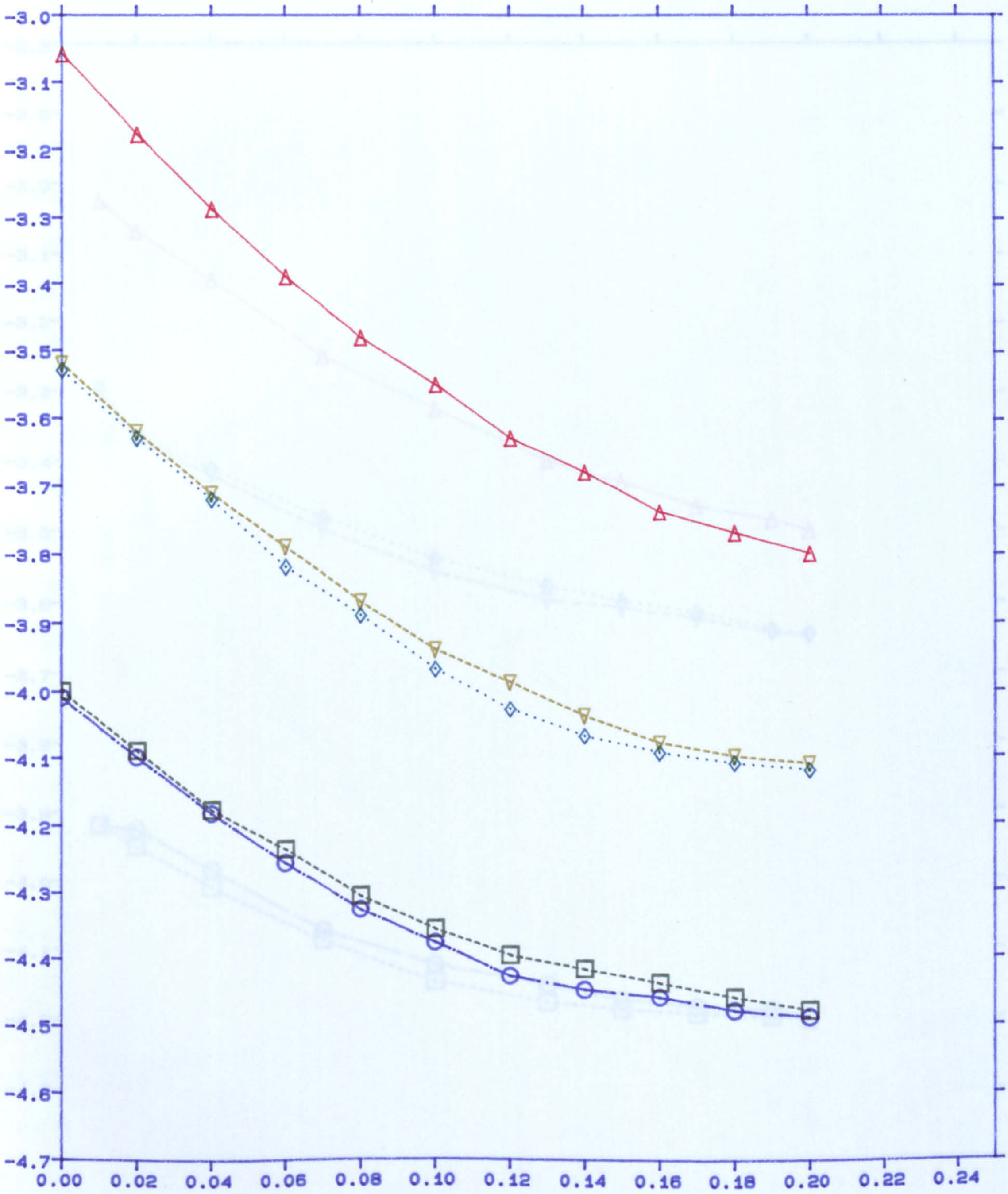


FIG. F8.11. GGRAN1PC CONFINING PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITIC GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATIONS

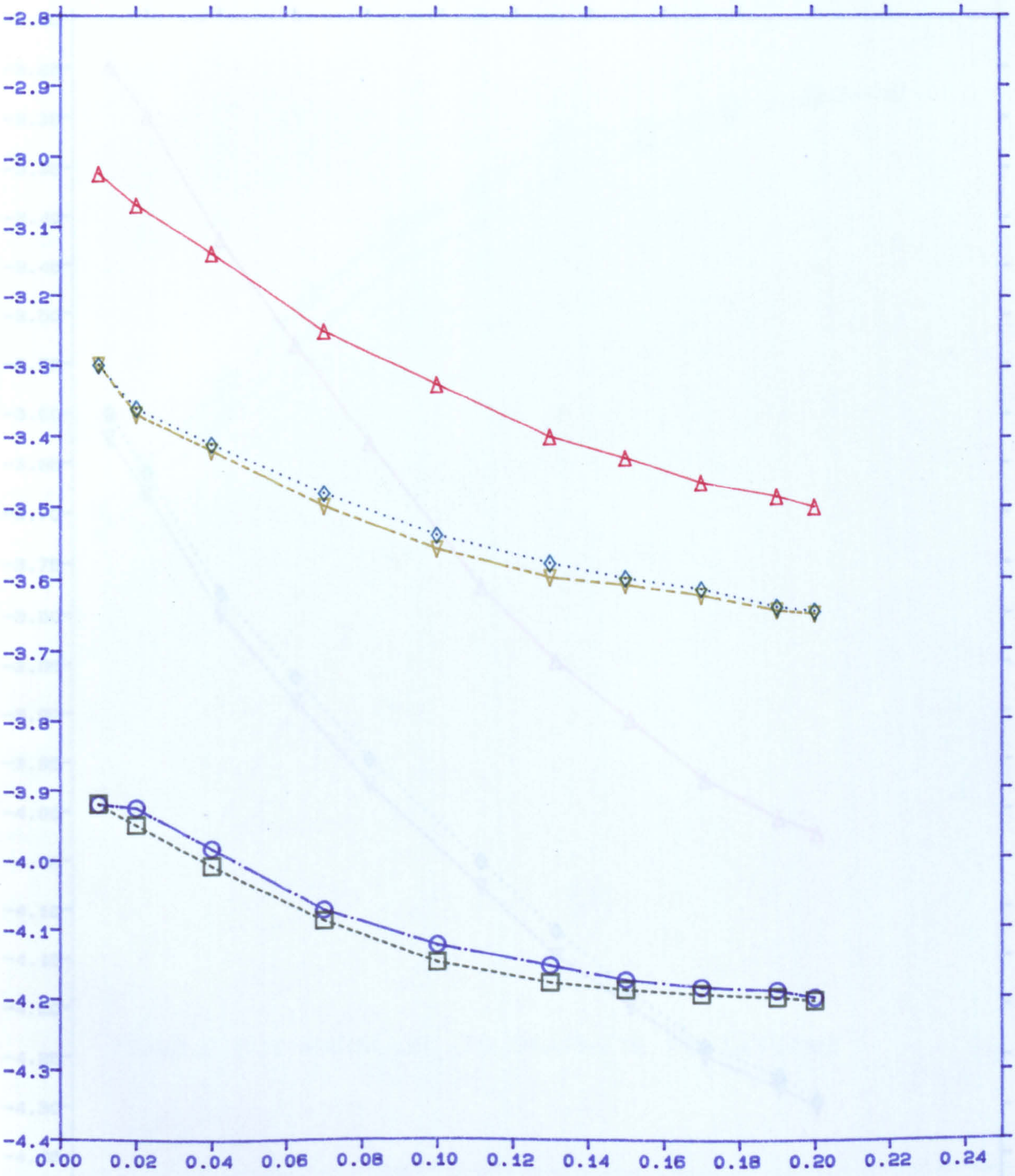


FIG. F8.12. GGRAN2PC CONFINING PRESSUREVARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITIC GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATIONS

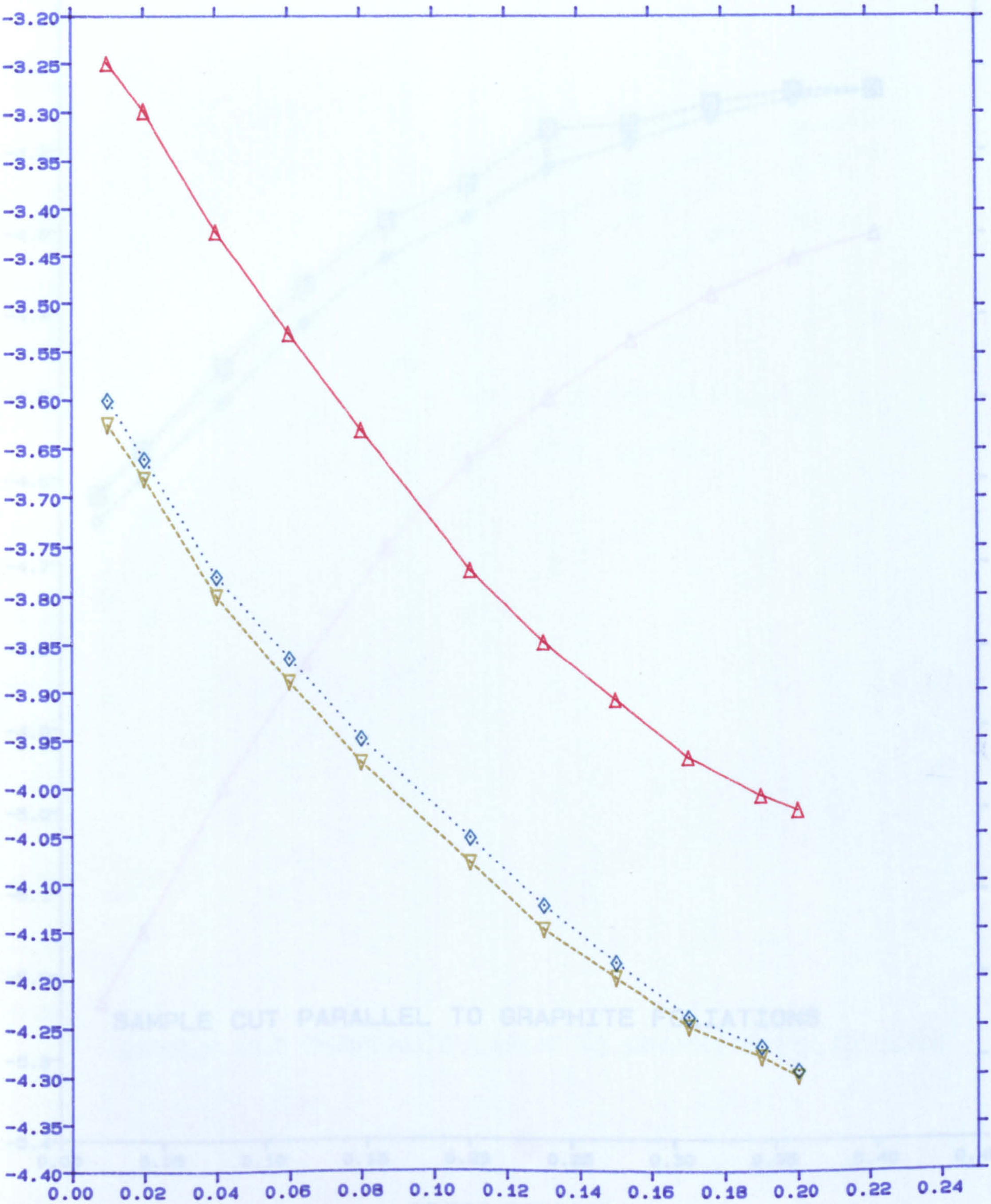


FIG. F8.13. GGRAN3PC CONFINING PRESSURE VARIATION OF DRY GRAPHITE GARNET GRANULITE

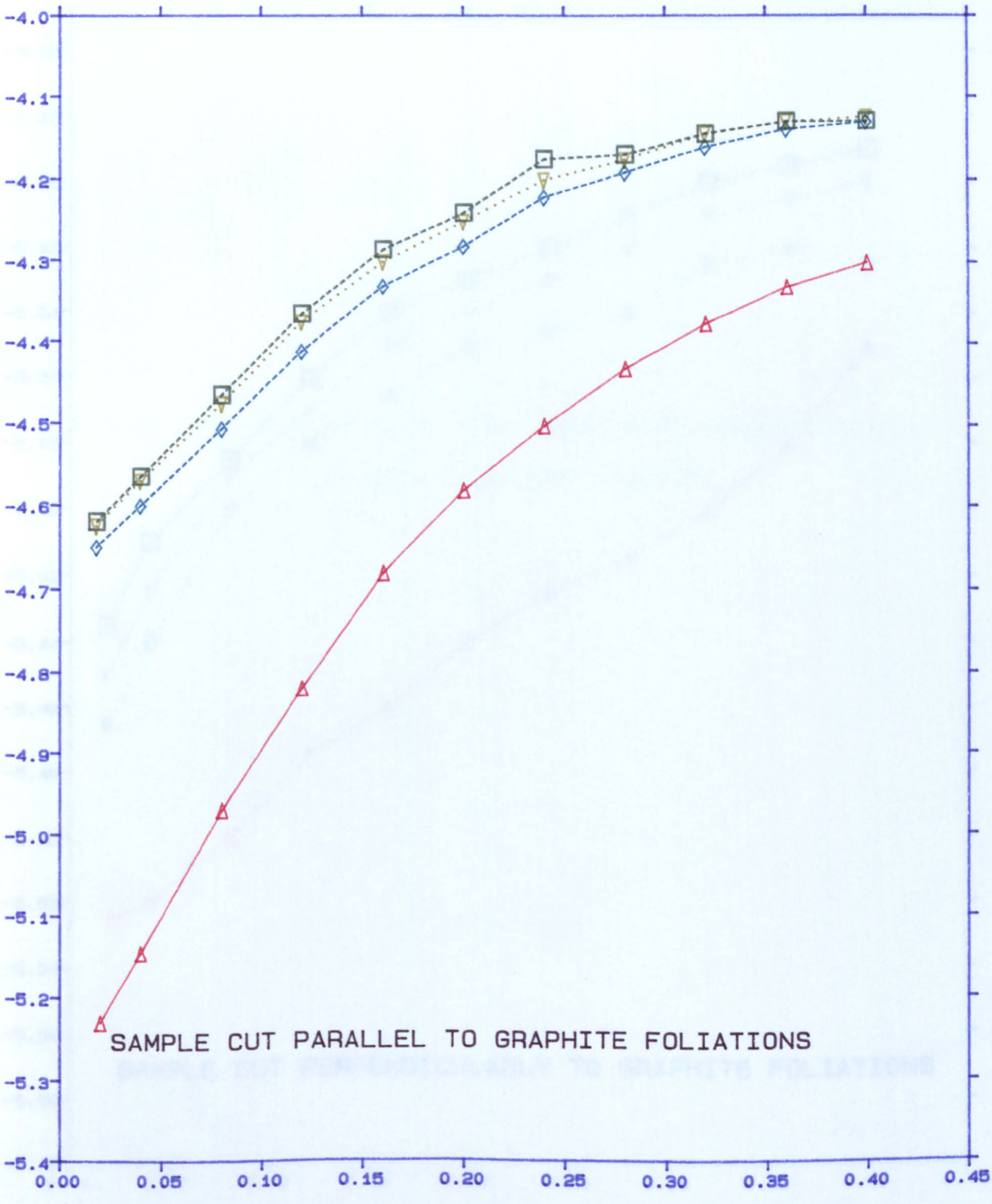
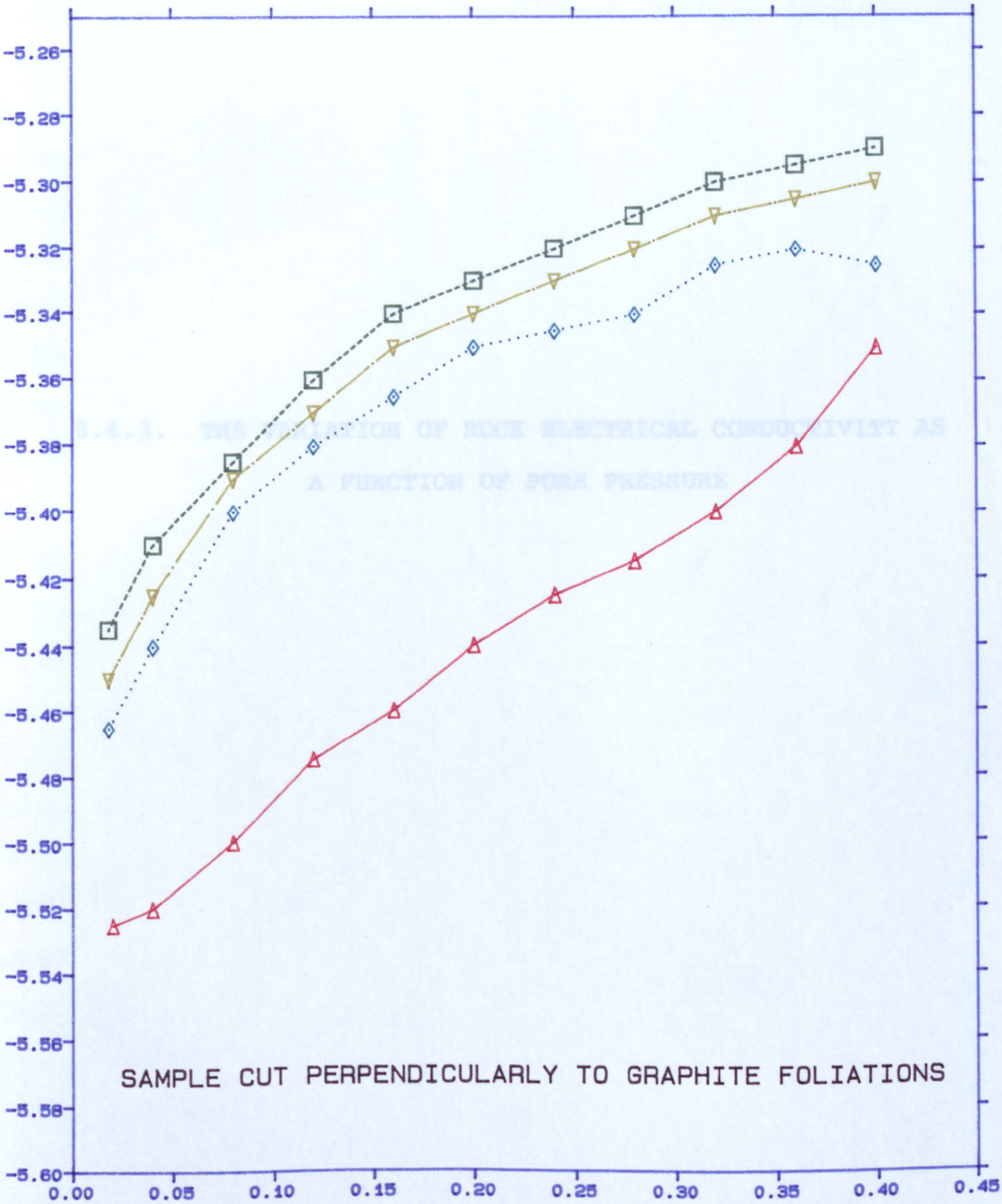


FIG. F8.14. GGRAN4PC CONFINING PRESSURE VARIATION OF A DRY GRAPHITE GARNET GRANULITE



8.4.3. THE VARIATION OF ROCK ELECTRICAL CONDUCTIVITY AS A FUNCTION OF PORE PRESSURE

FIG F8.15. IVD1PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

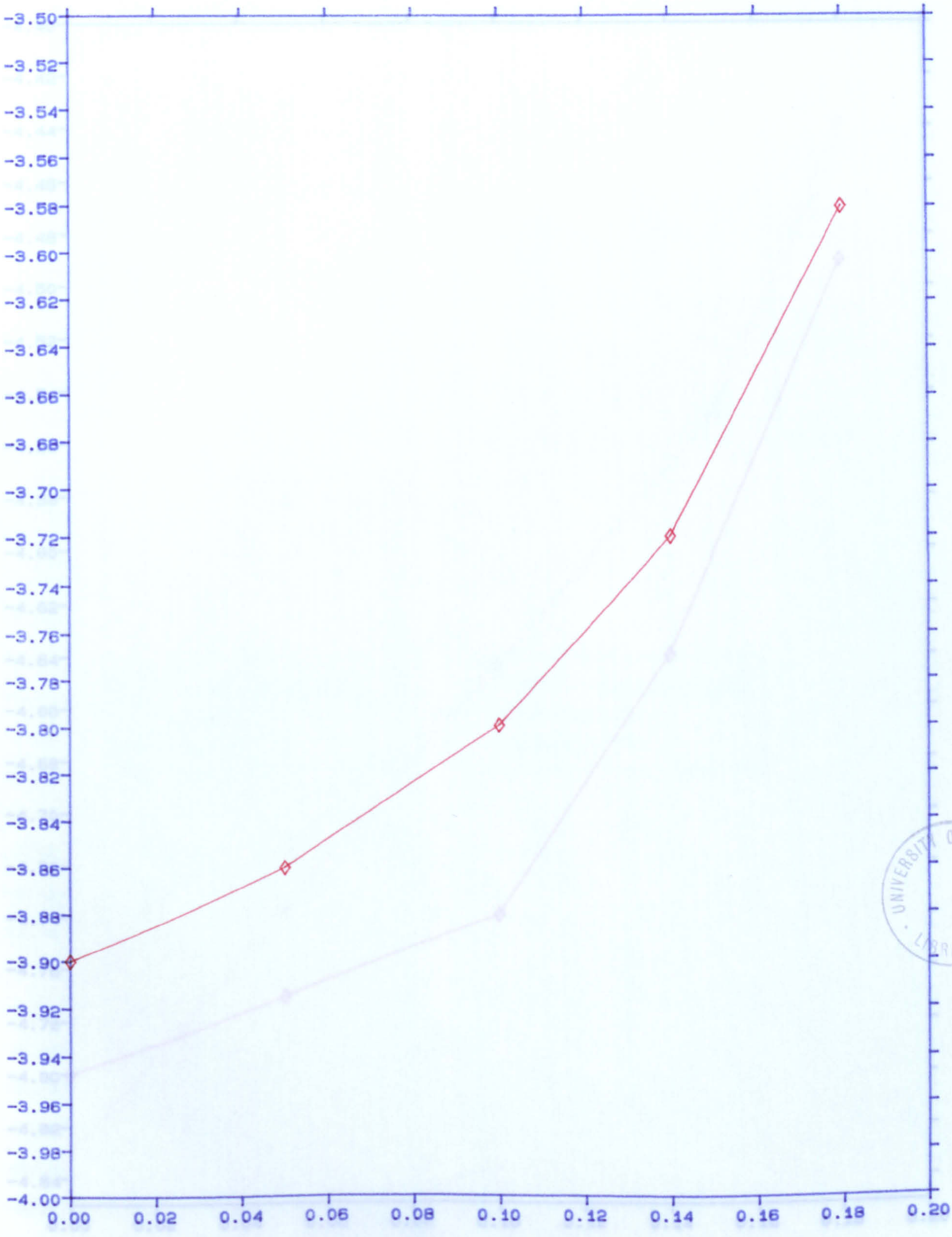


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FIG. F8.16. IVD2PP PORE PRESSURE VARIATION FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

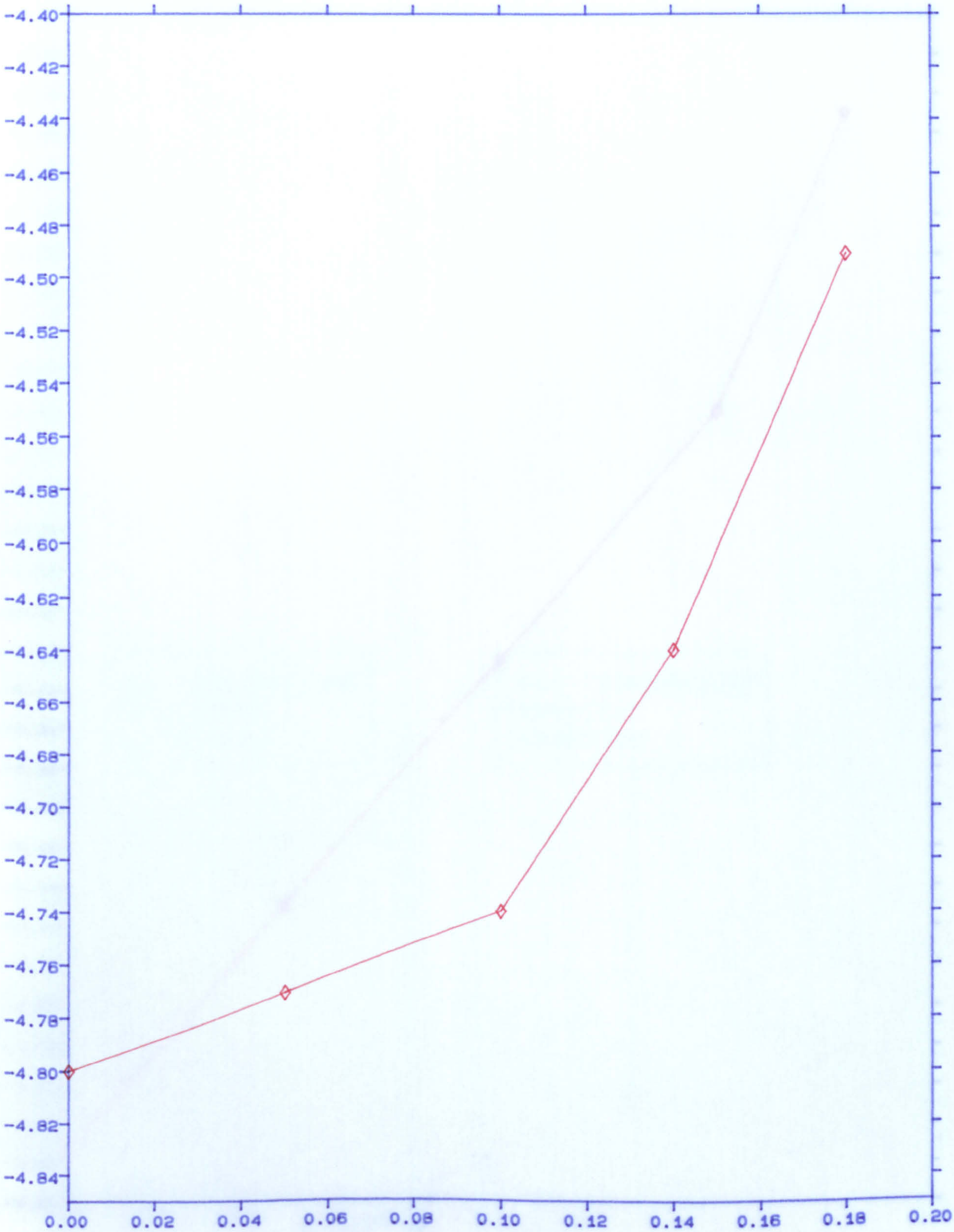


FIG. F8.17. IVE1PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

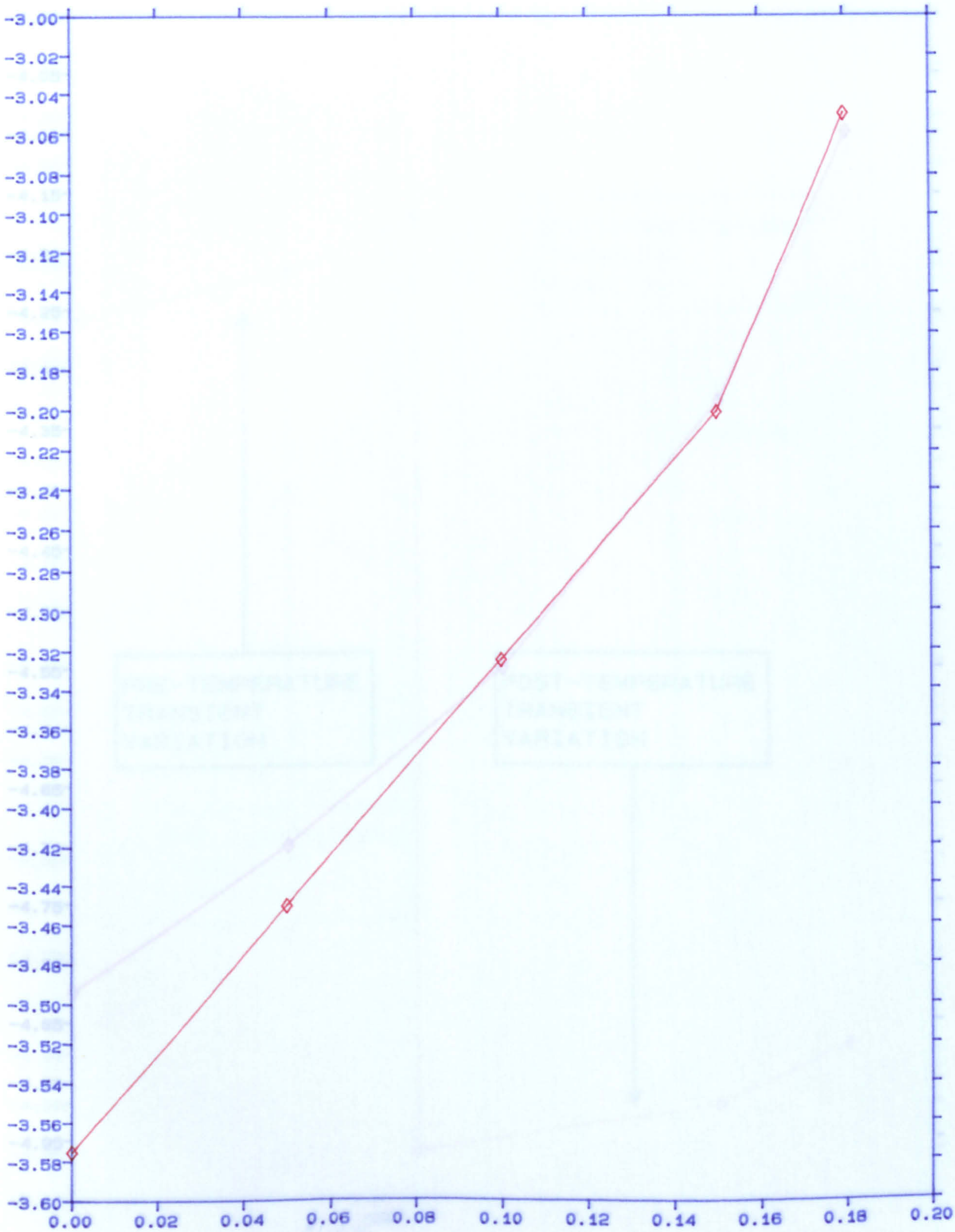


FIG. F8.18. IVE2PP PORE PRESSURE VARIATION OF DISTILLED WATER SATURATED BASIC GRANULITE

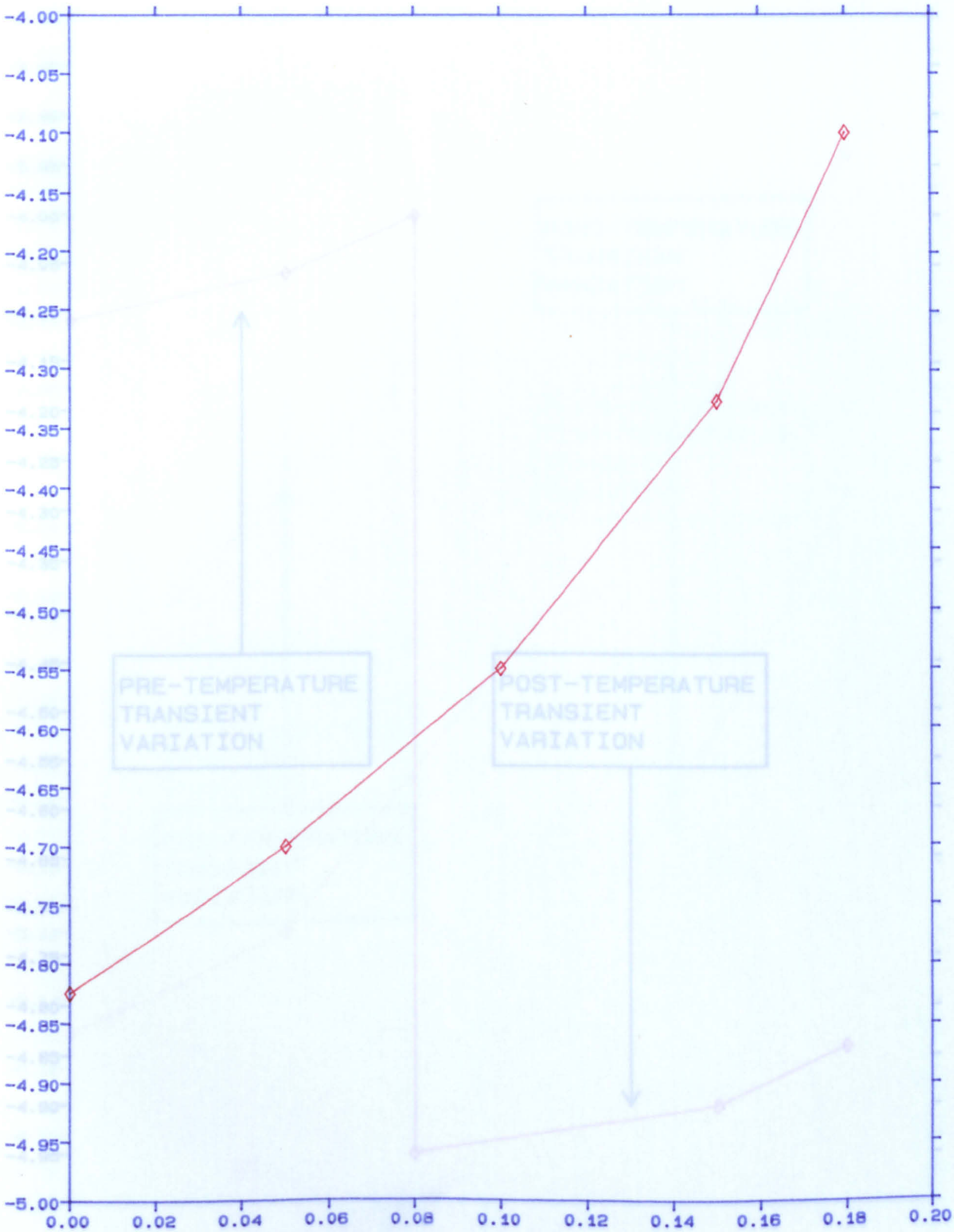


FIG. F8.19. MER21PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE

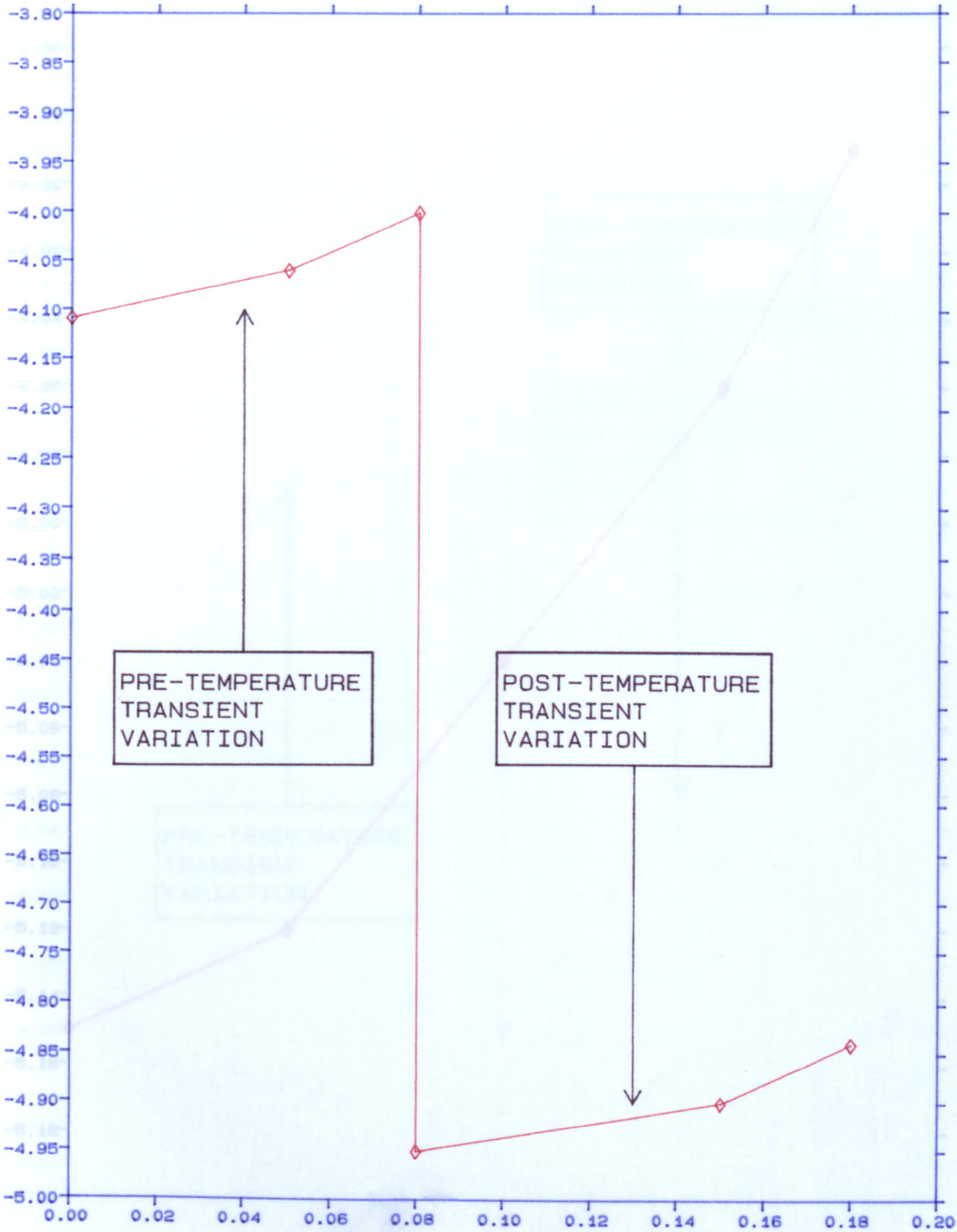


FIG. F8.20. MER22PP PORE PRESSURE VARIATION OF DISTILLED WATER SATURATED ORTHOPYROXENITE

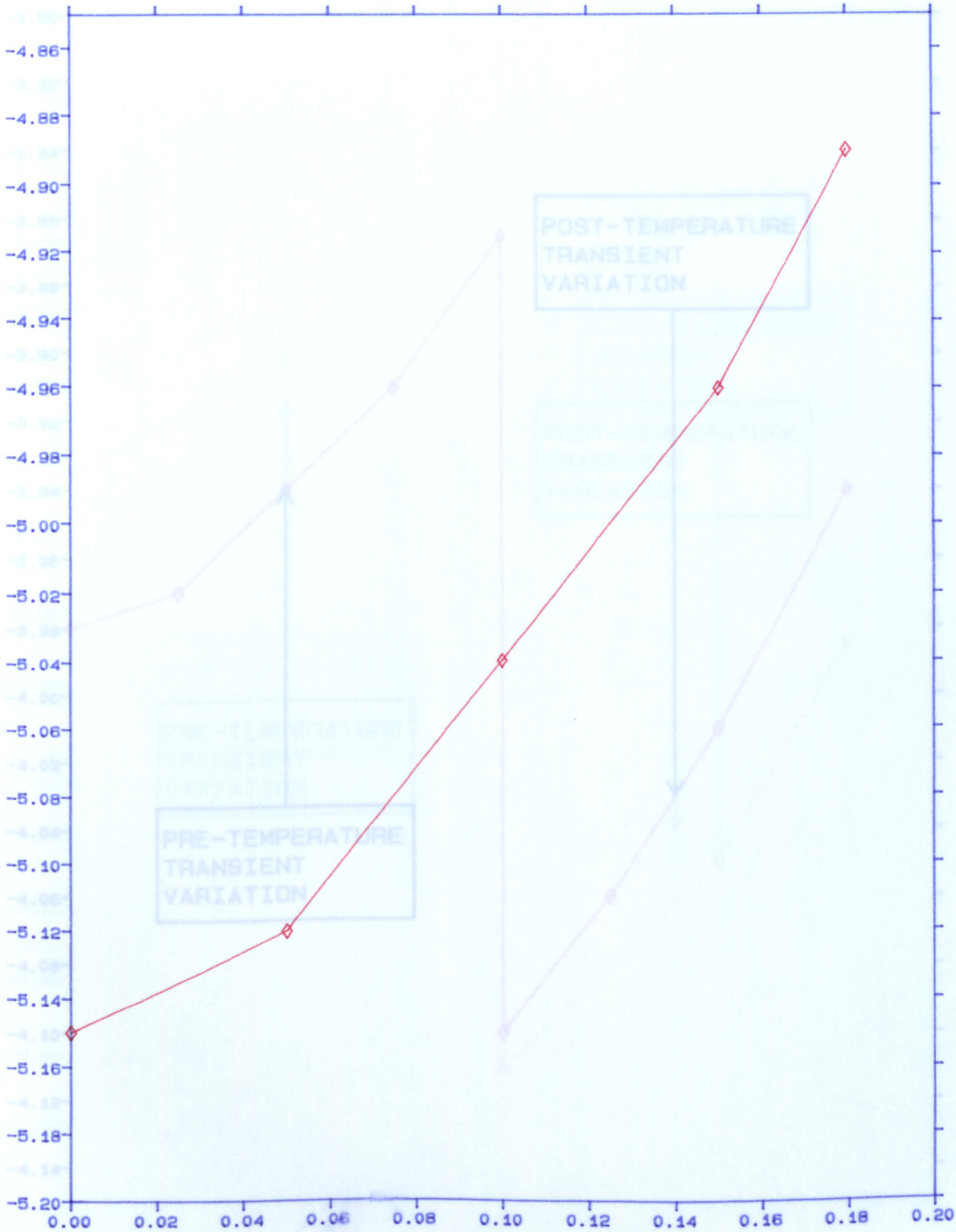


FIG. F8.21. A8BPP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

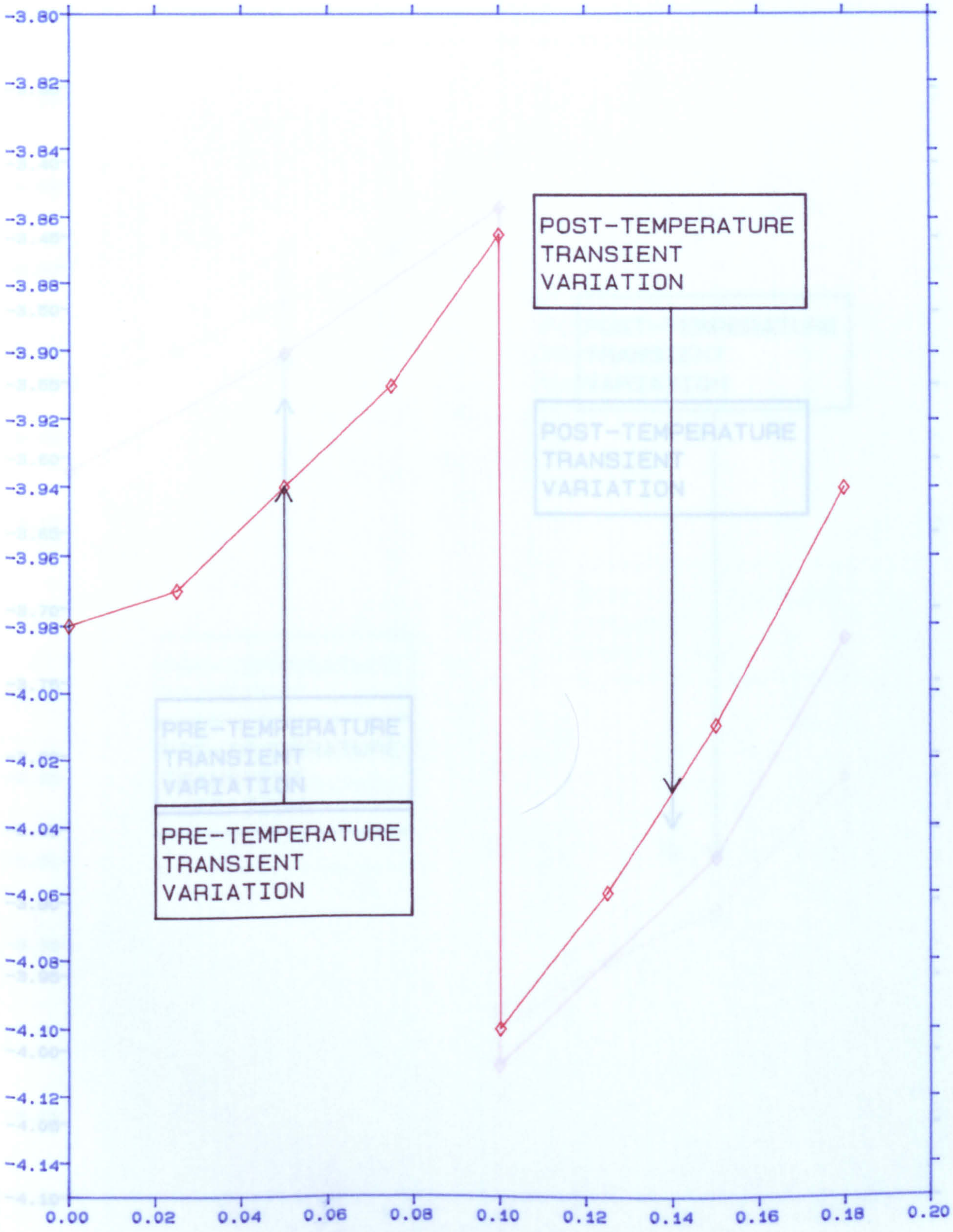


FIG. F8.22. L33BPP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED GRANODIORITE

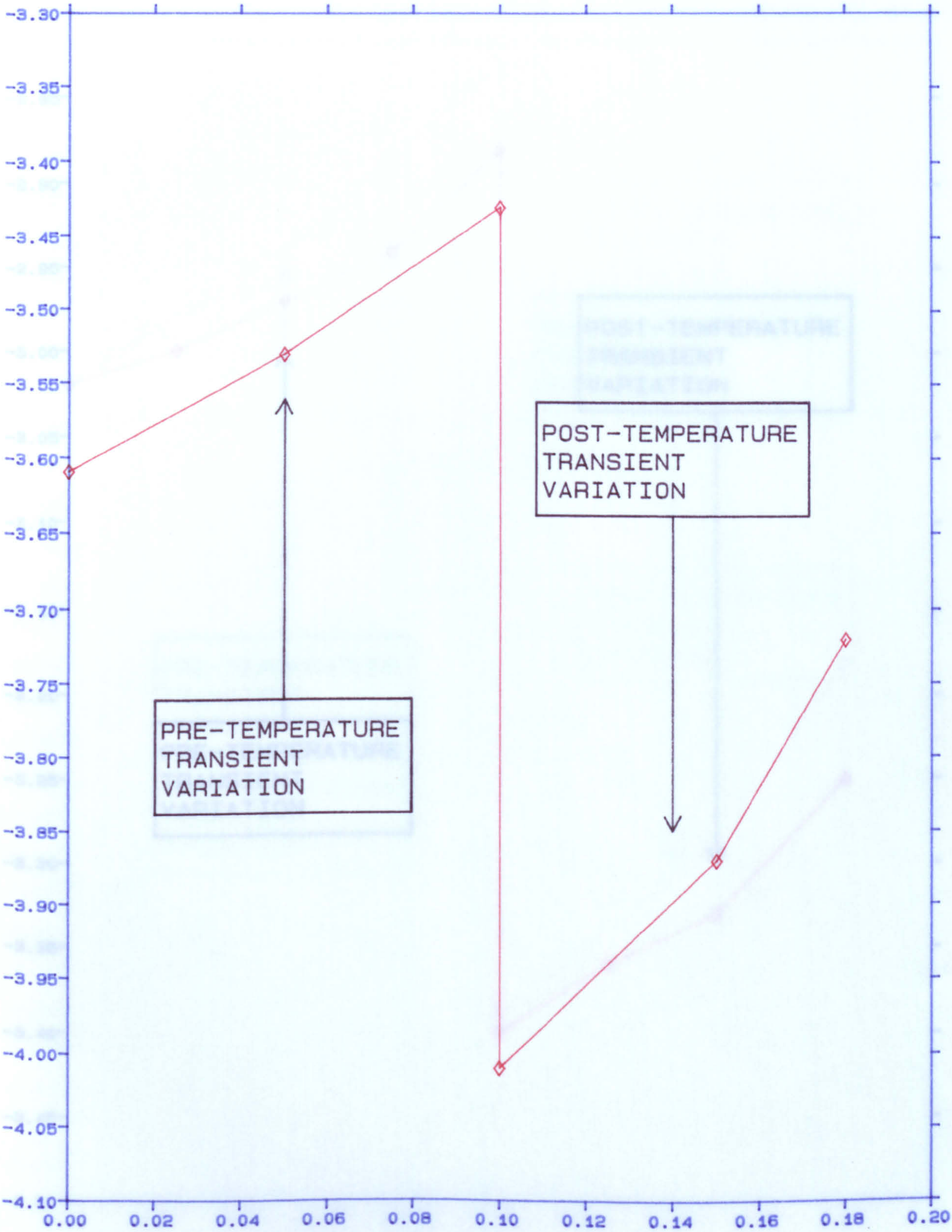


FIG. F8.23. L32PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

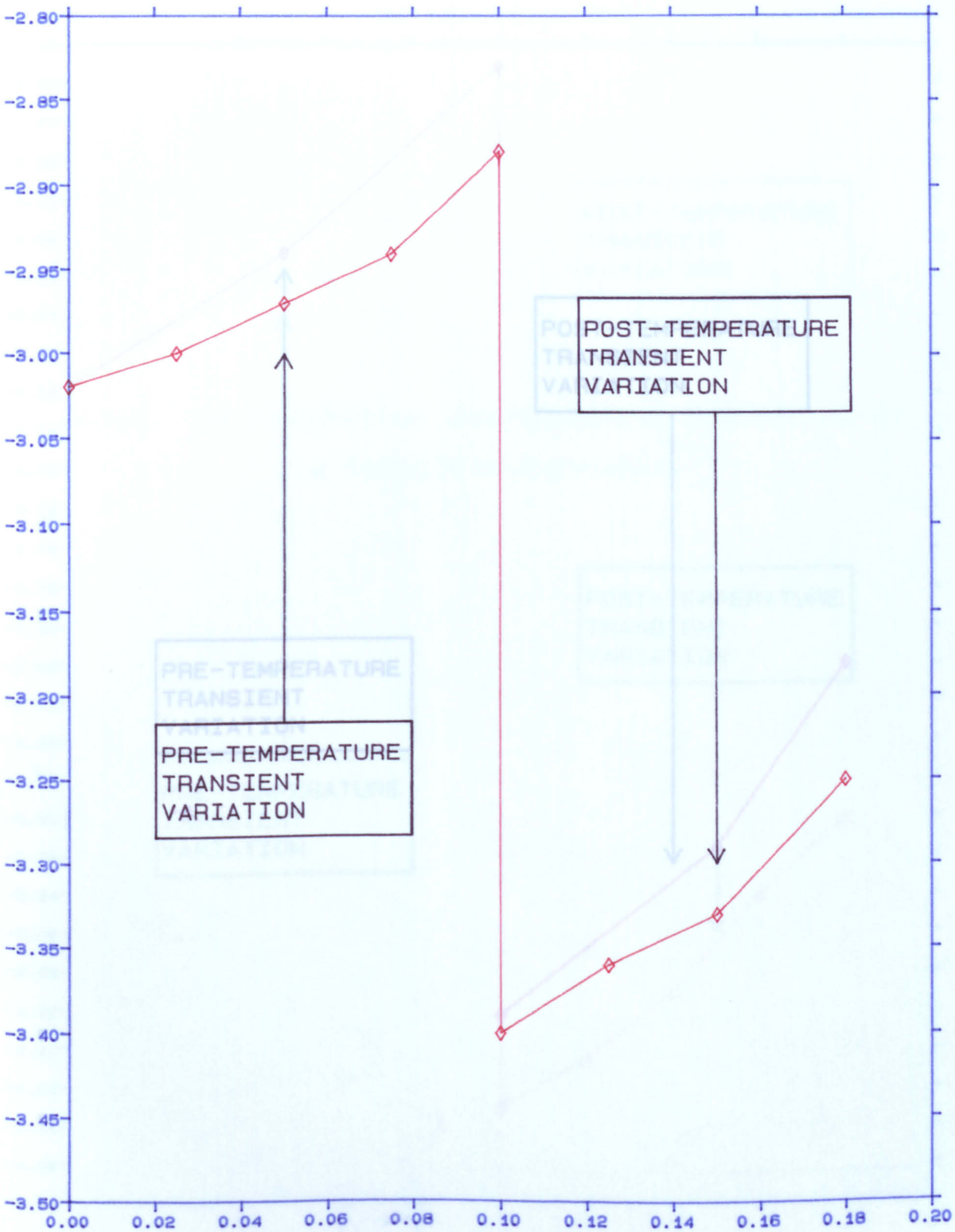


FIG. F8.24. V1PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED GRANITE

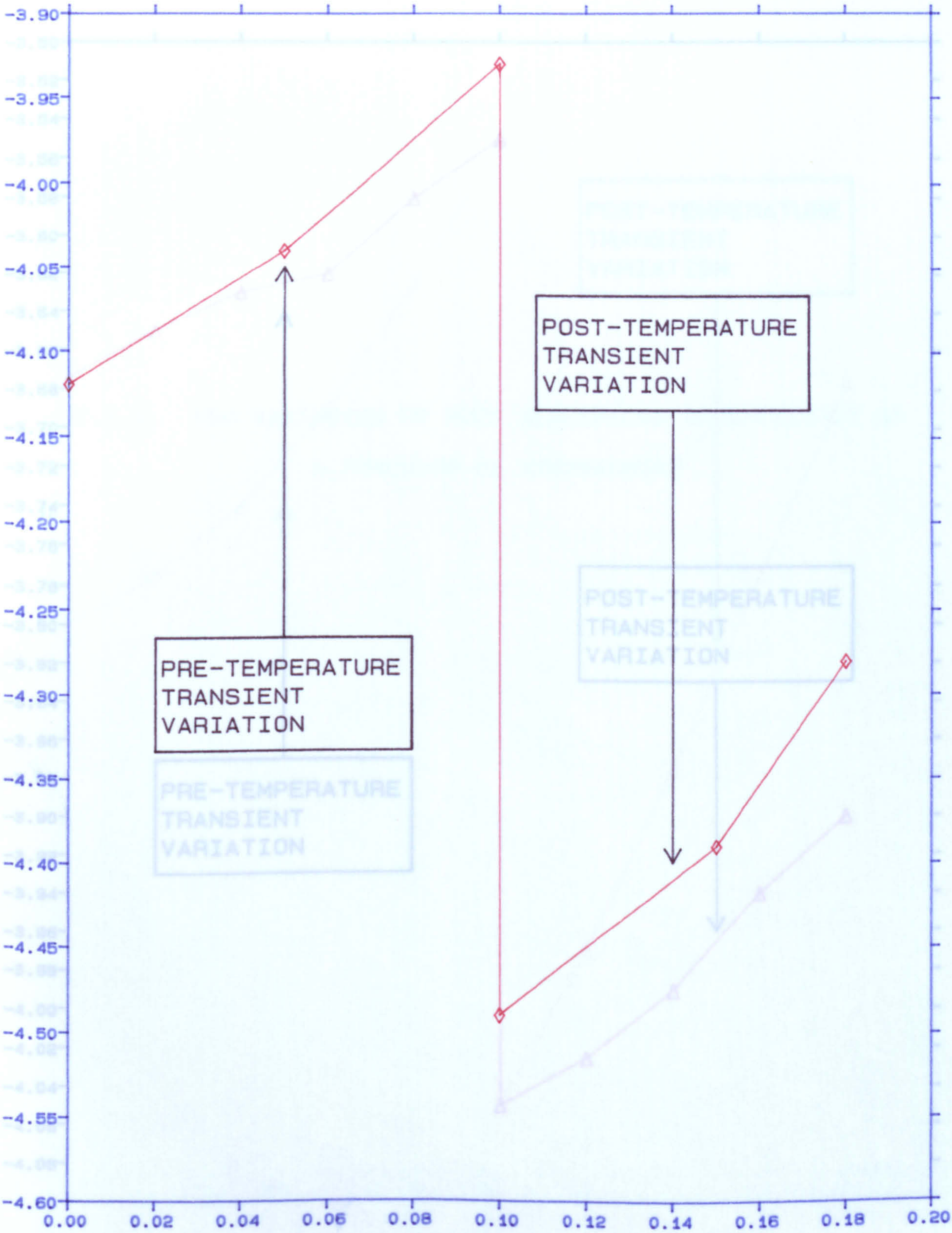


FIG. F8.25. GGRAN1PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED GRAPHITIC GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATIONS

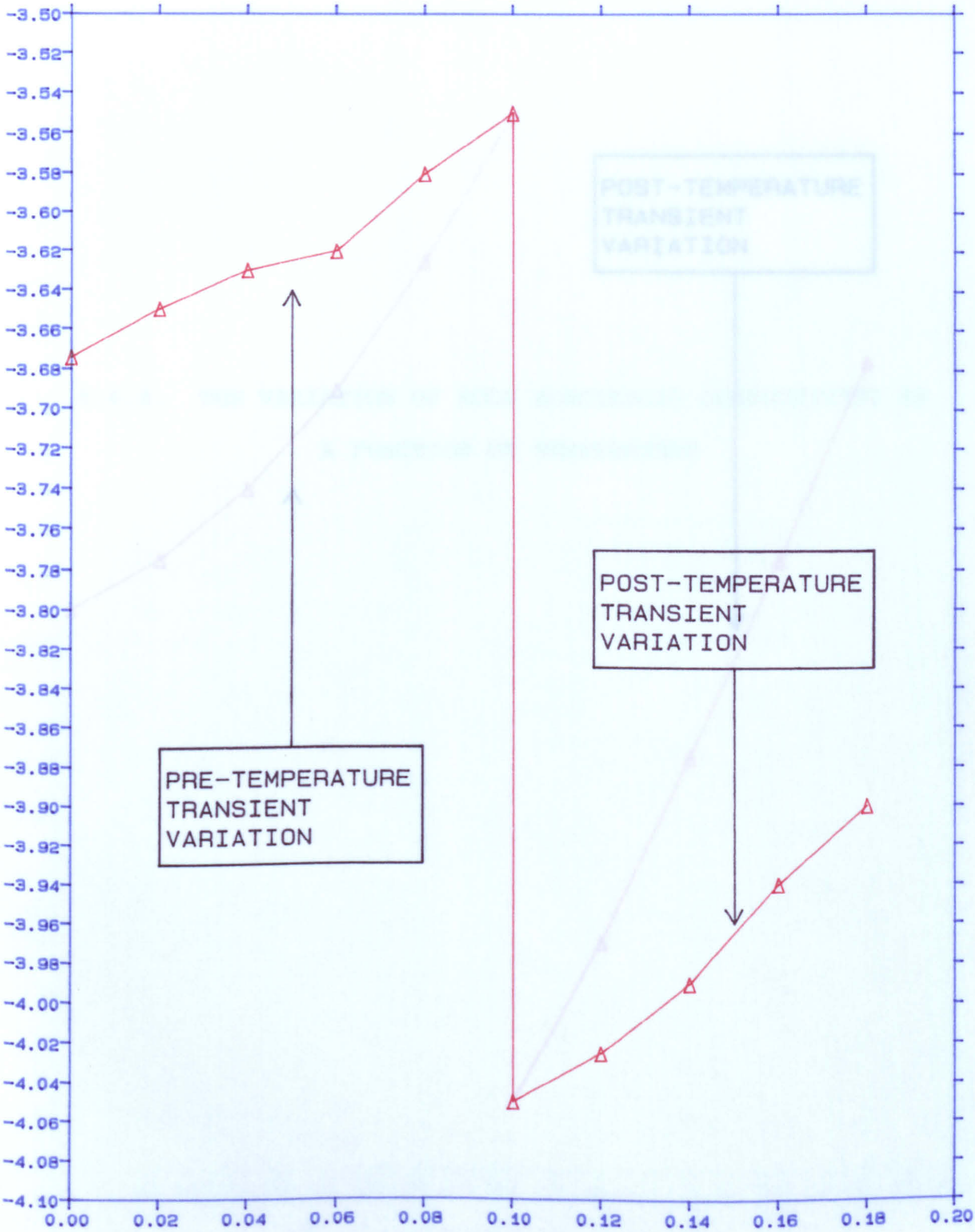
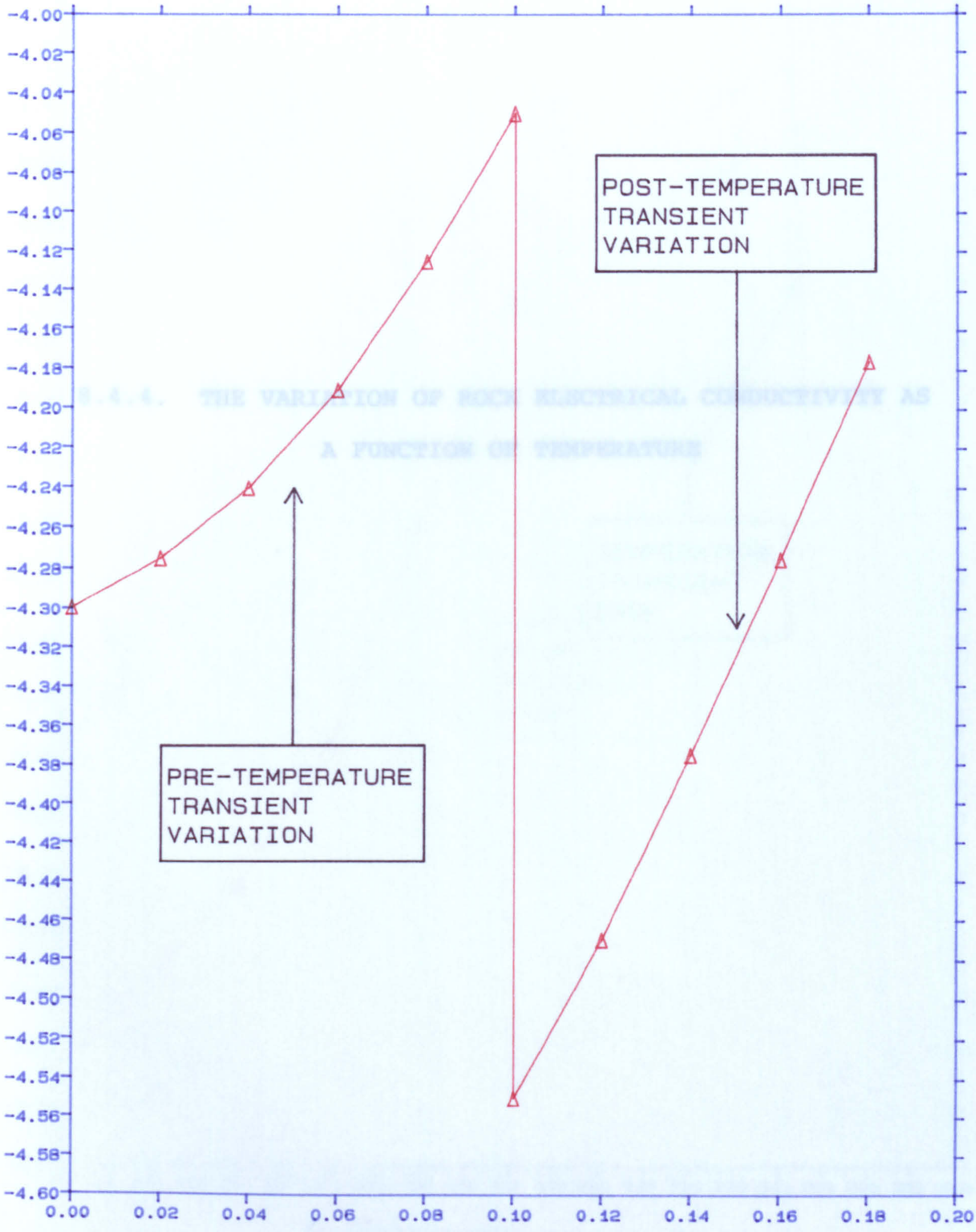


FIG. F8.26. GGRAN2PP PORE PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED GRAPHITIC GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATIONS



8.4.4. THE VARIATION OF ROCK ELECTRICAL CONDUCTIVITY AS A FUNCTION OF TEMPERATURE

FIG. F8.27. IVD1TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

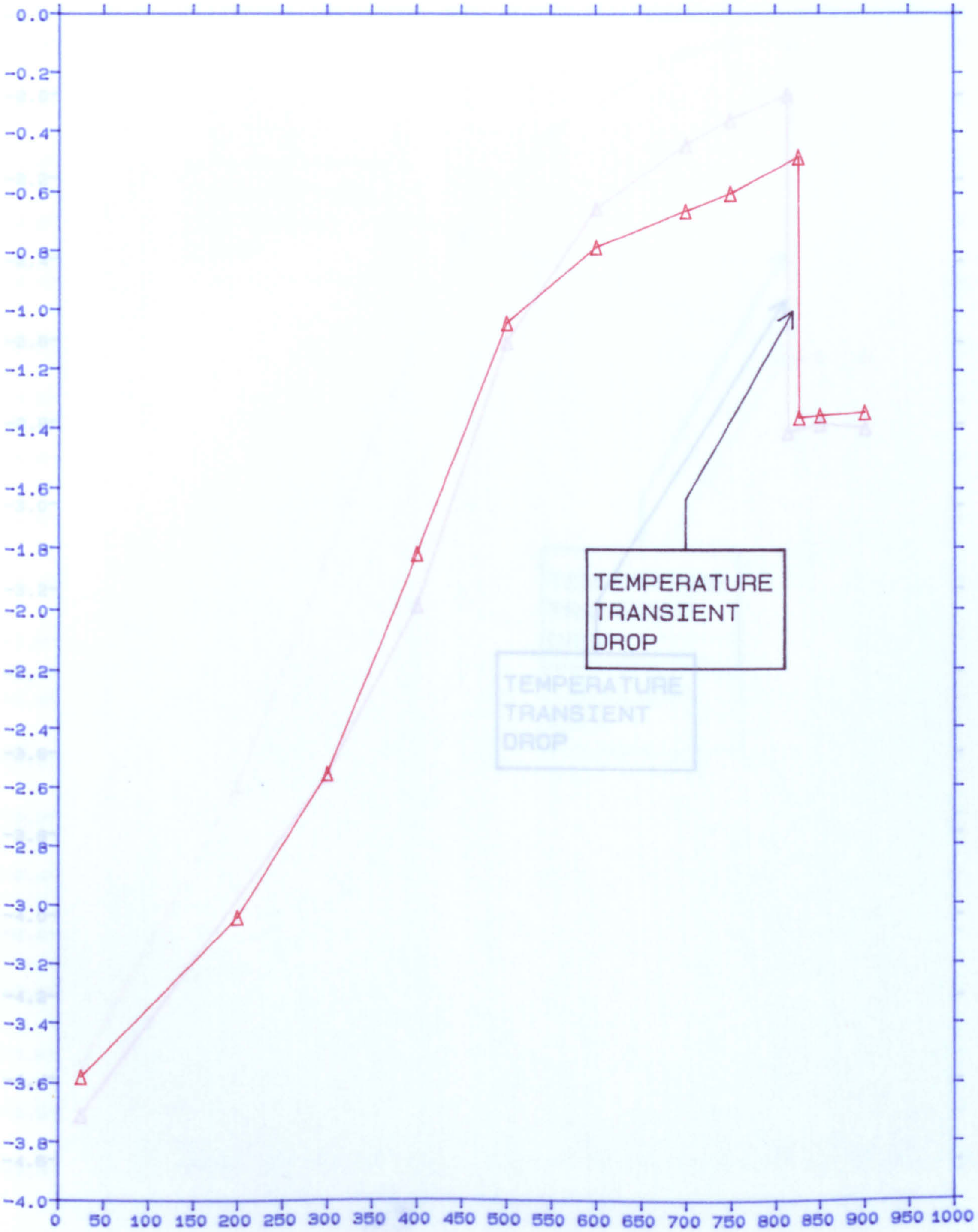


FIG. F8.28. IVD2TT INITIAL TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

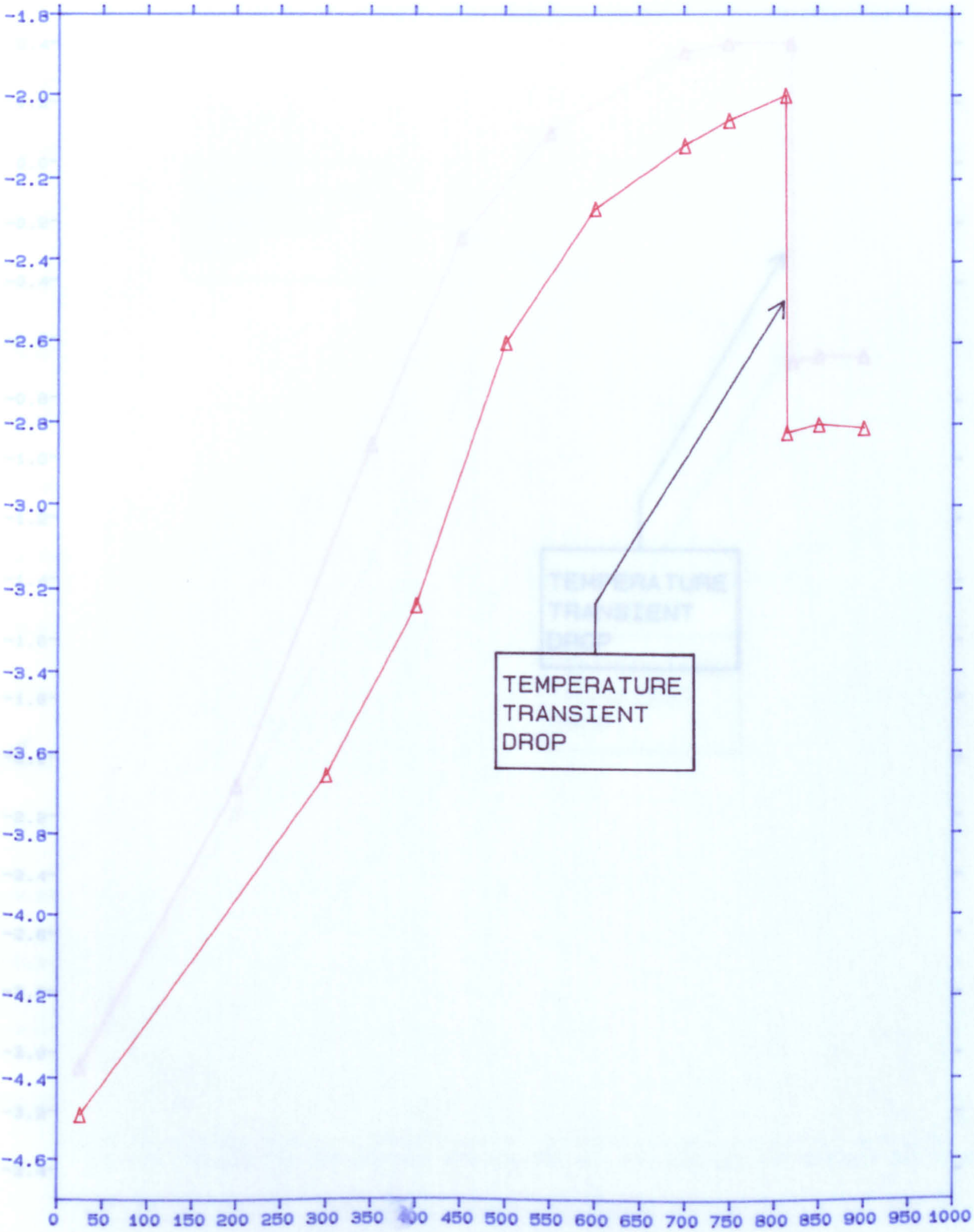


FIG. F8.29. IVE1TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

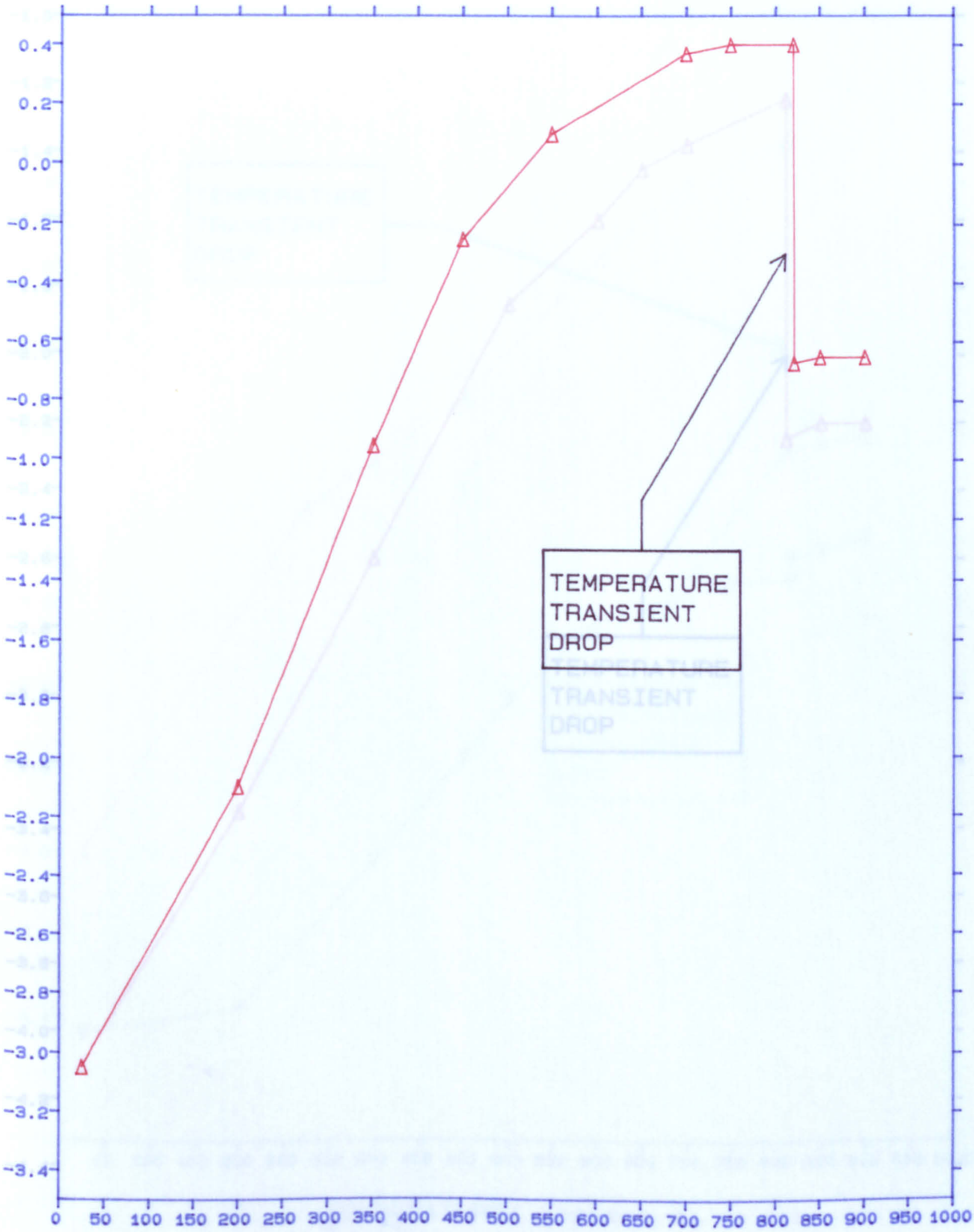


FIG. F8.30. IVE2TT INITIAL TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED BASIC GRANULITE

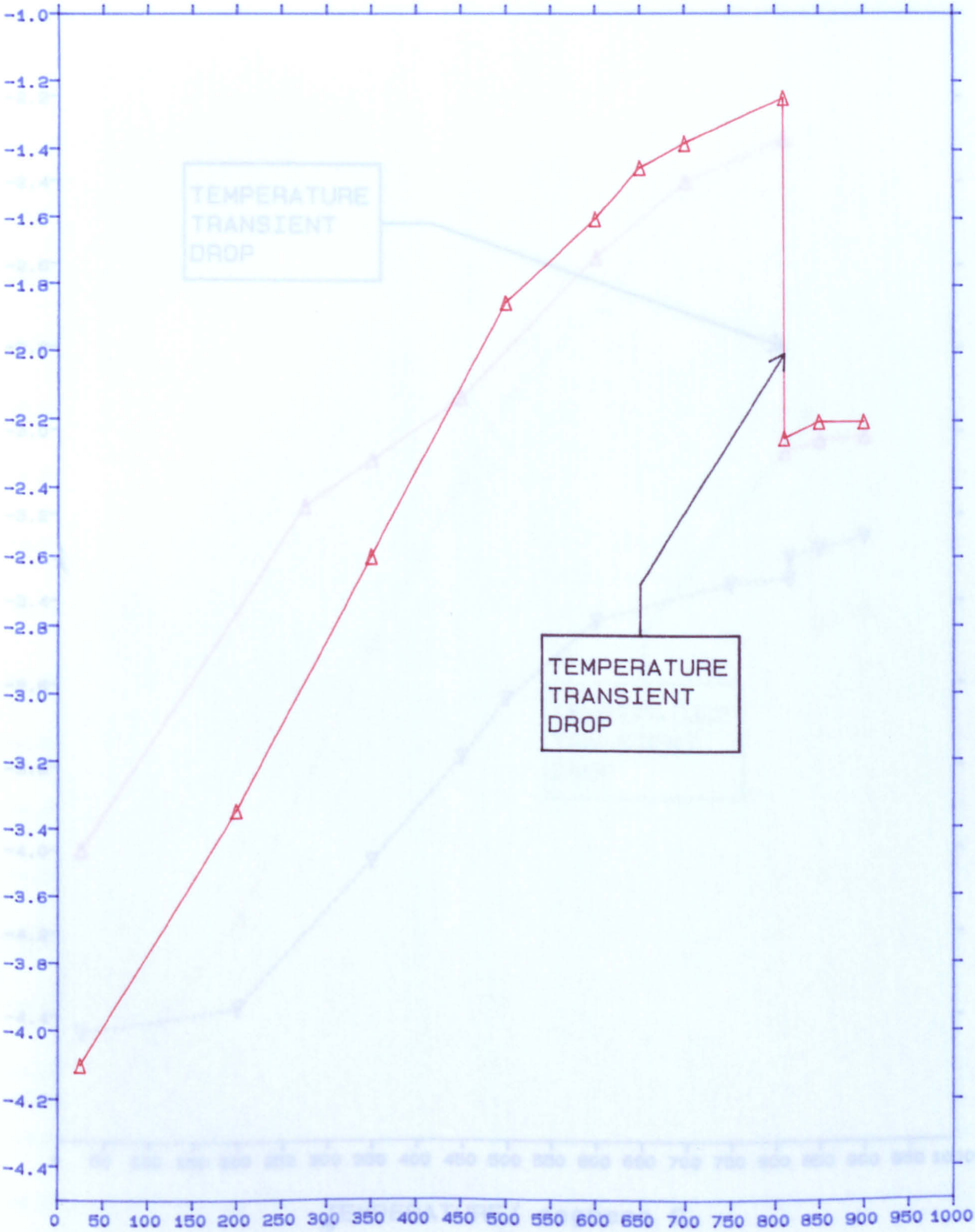
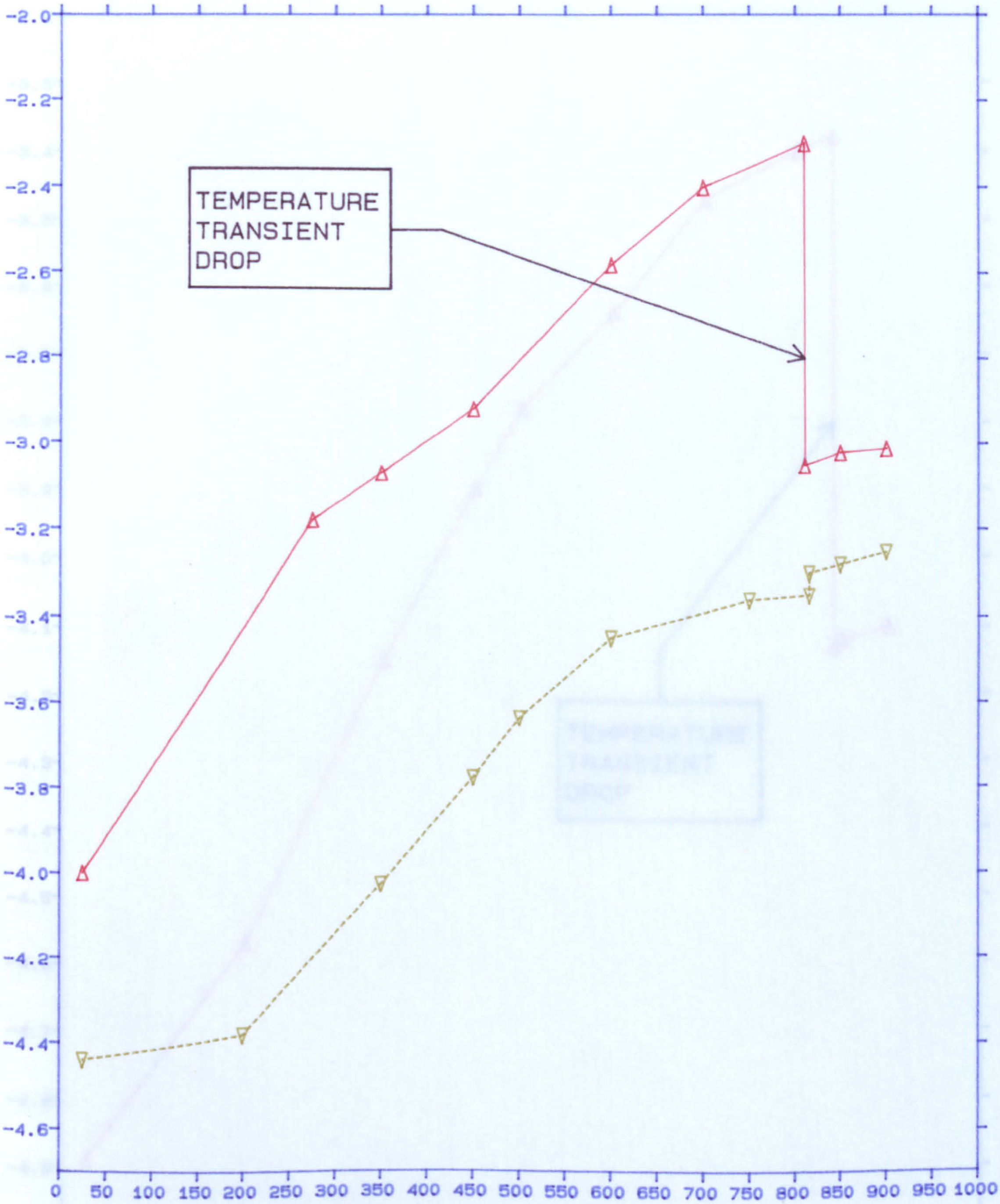


FIG. F8.31. MER21TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE



—Δ— PRE-TEMPERATURE TRANSIENT VARIATION AT A PORE PRESSURE OF 0.1 GPa

-----▽----- SUBSEQUENT INITIAL TEMPERATURE VARIATION AT A PORE PRESSURE OF 0.18 GPa

FIG. F8.32. MER22TT INITIAL TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED ORTHOPYROXENITE

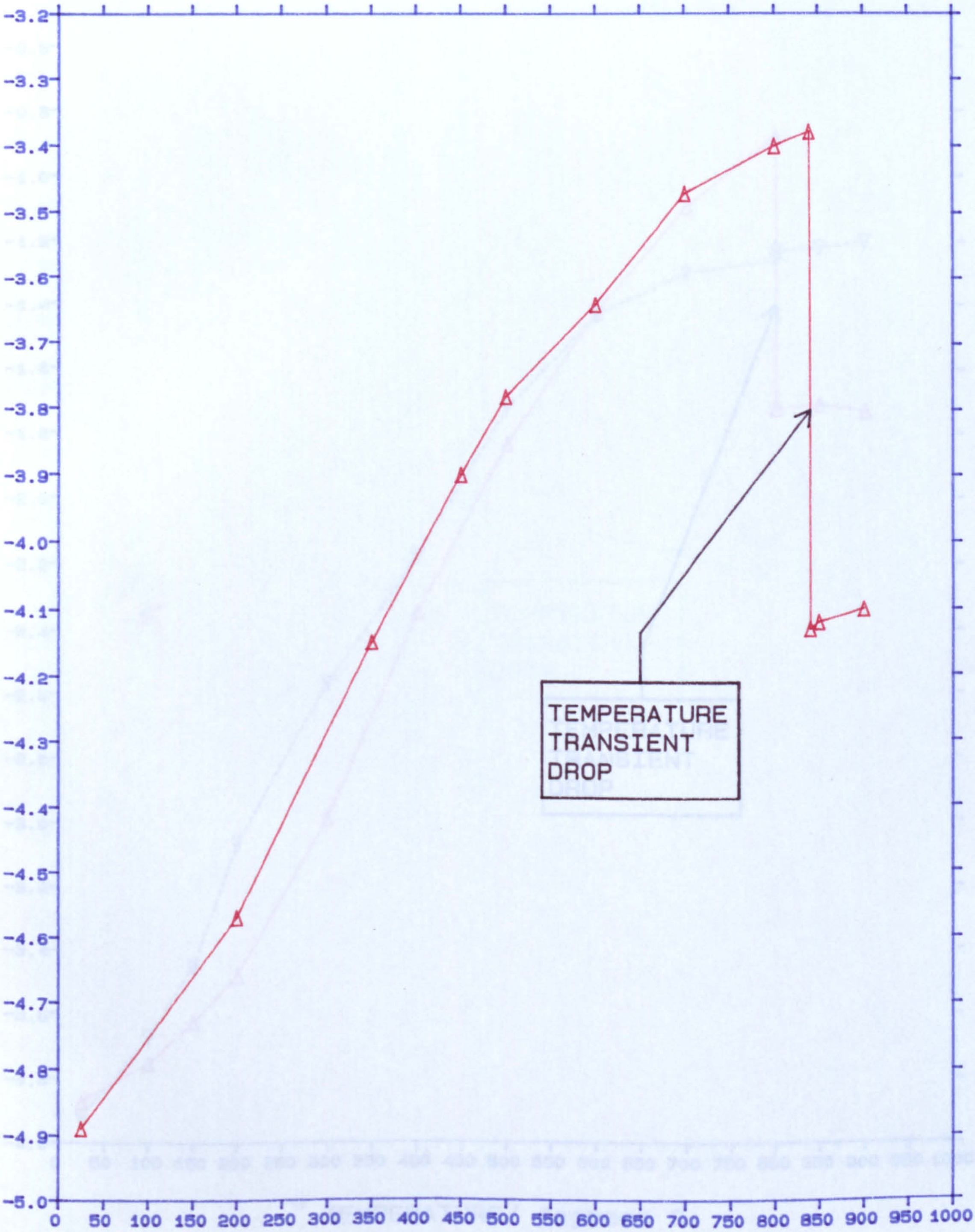


FIG. F8.33. A8BTT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

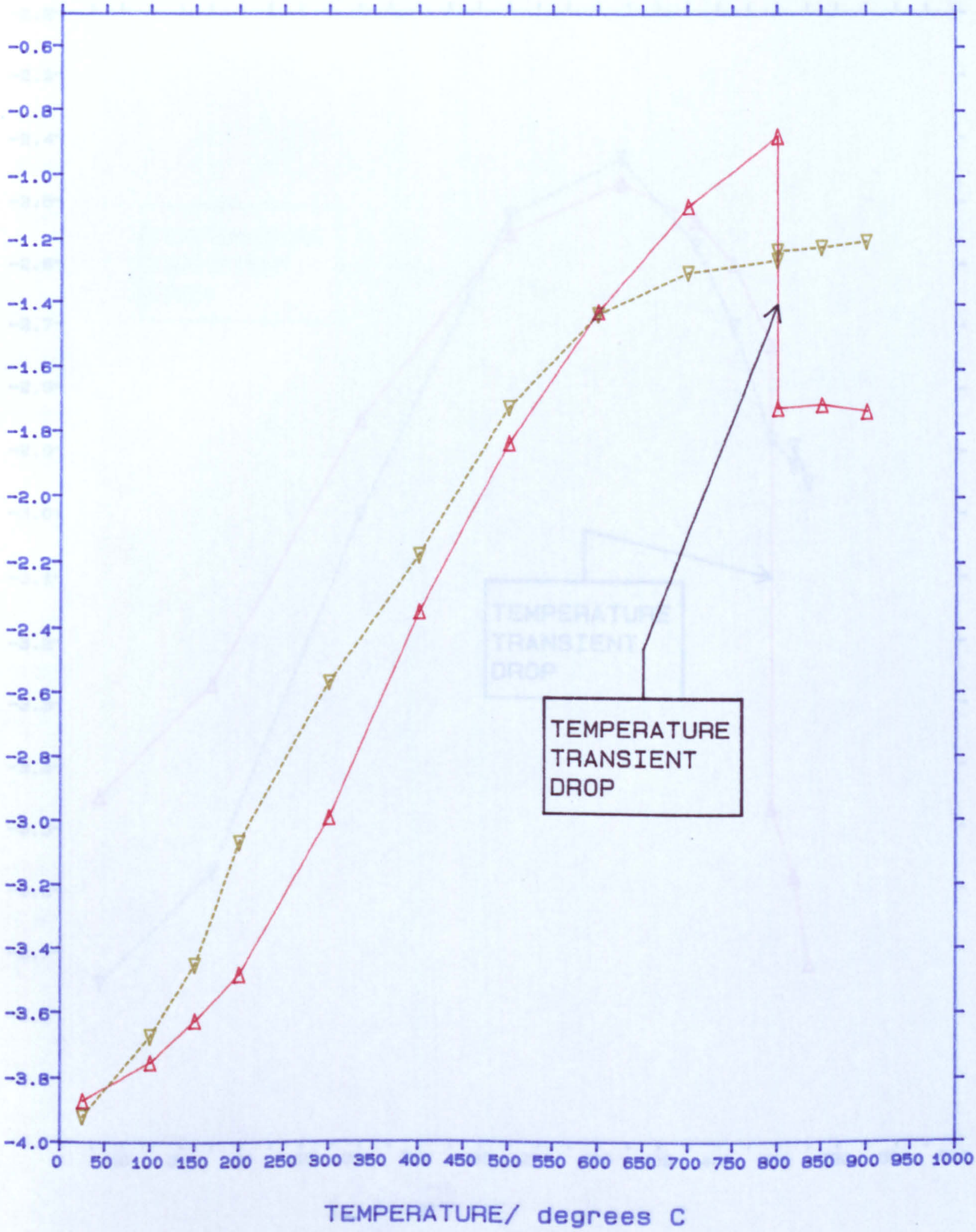
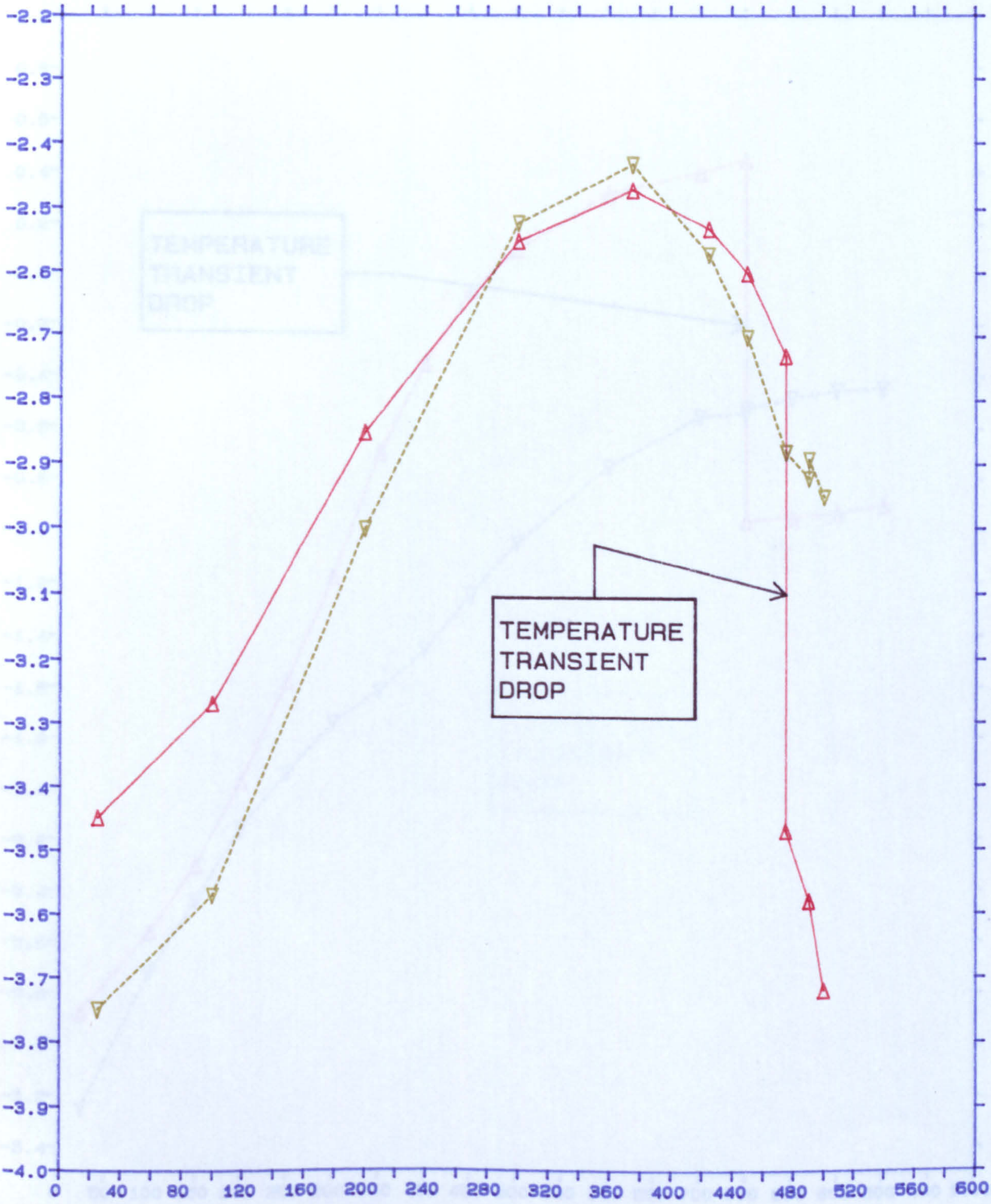


FIG. F8.34. L33BTT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE



—△— PRE-TEMPERATURE TRANSIENT VARIATION AT A PORE PRESSURE OF 0.1 GPa

----▽---- SUBSEQUENT INITIAL TEMPERATURE VARIATION AT A PORE PRESSURE OF 0.18 GPa

FIG. F8.35. L32TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

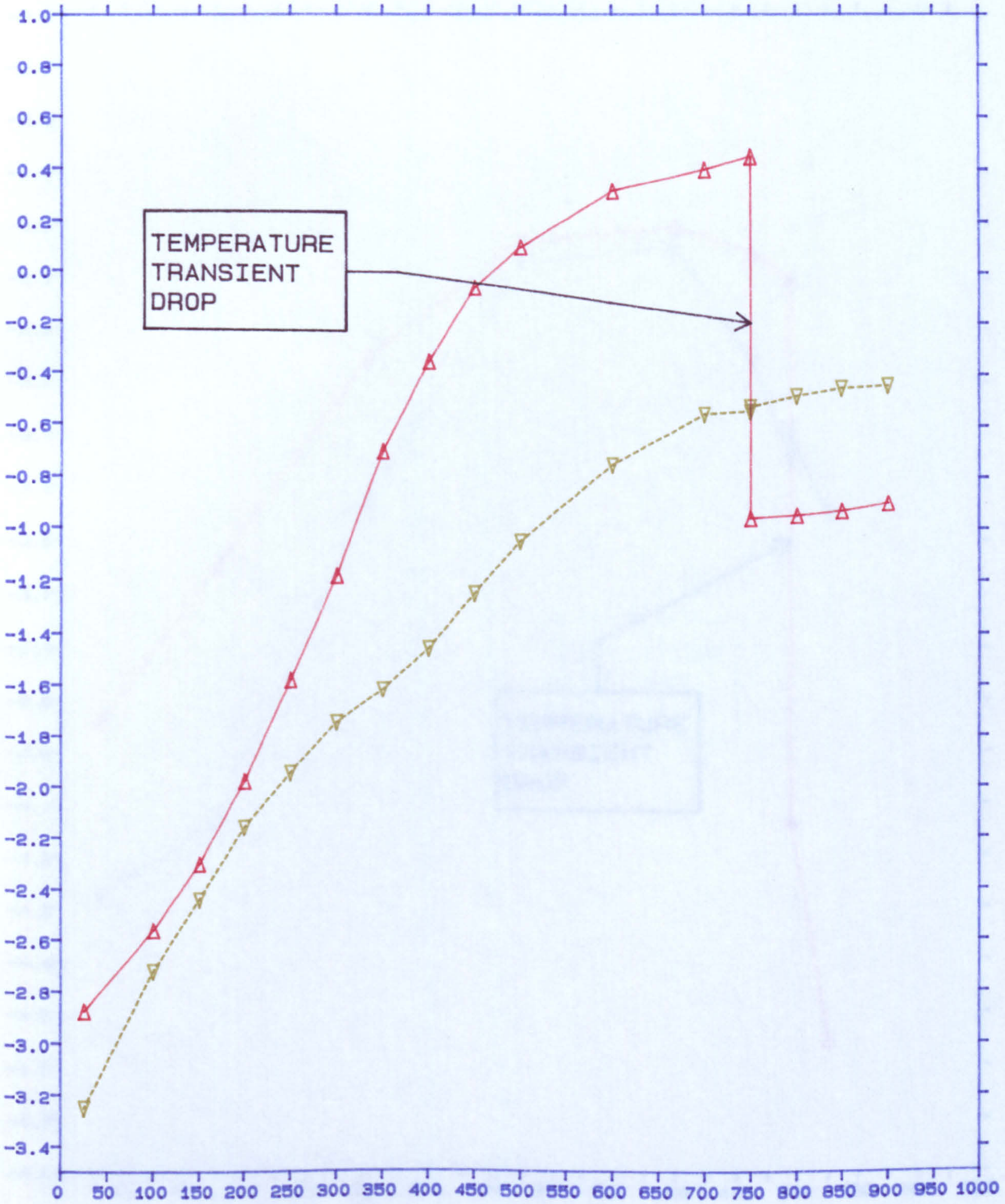


FIG. F8.36. V1TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANITE

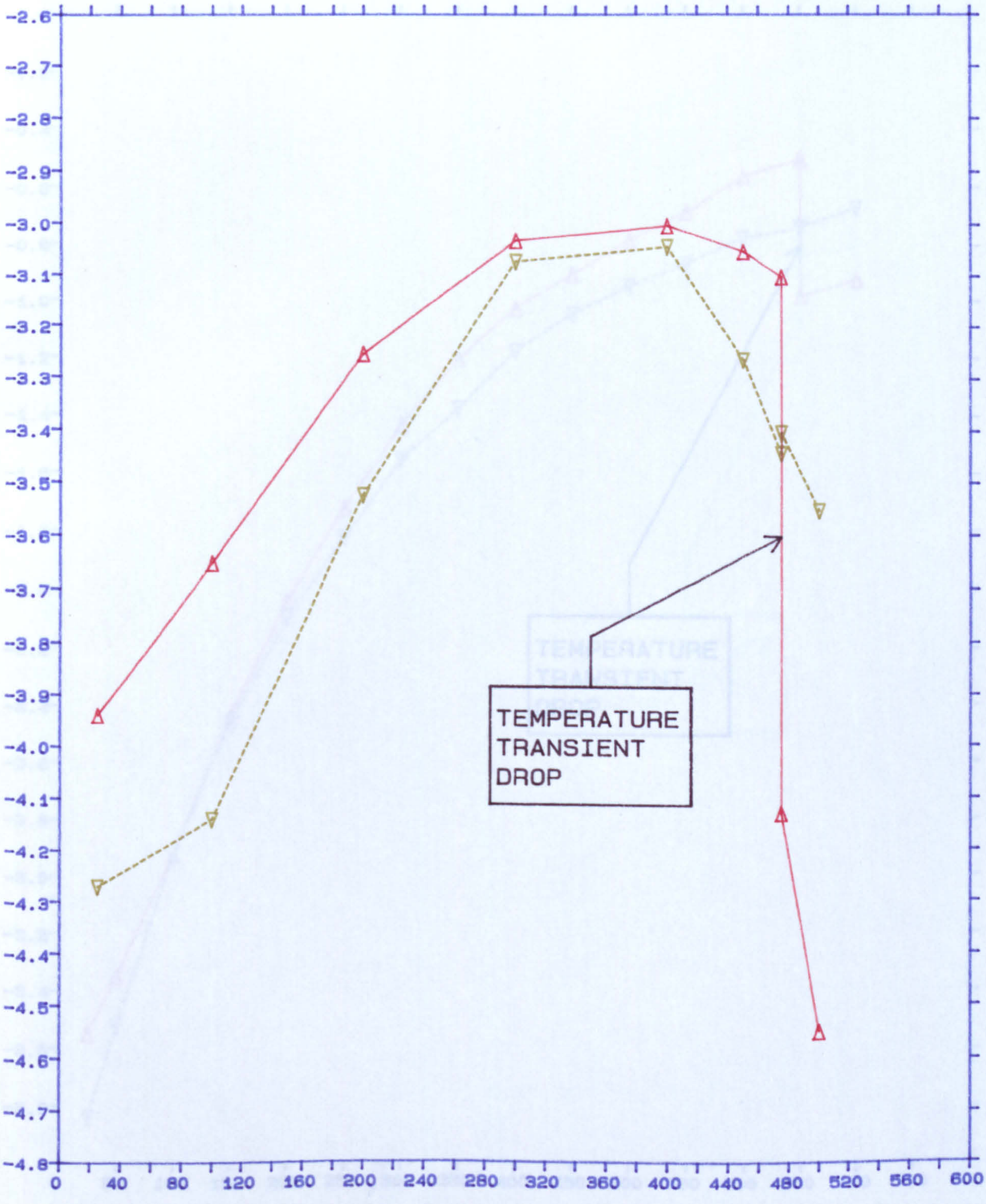


FIG. F8.37. GGRAN1TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE

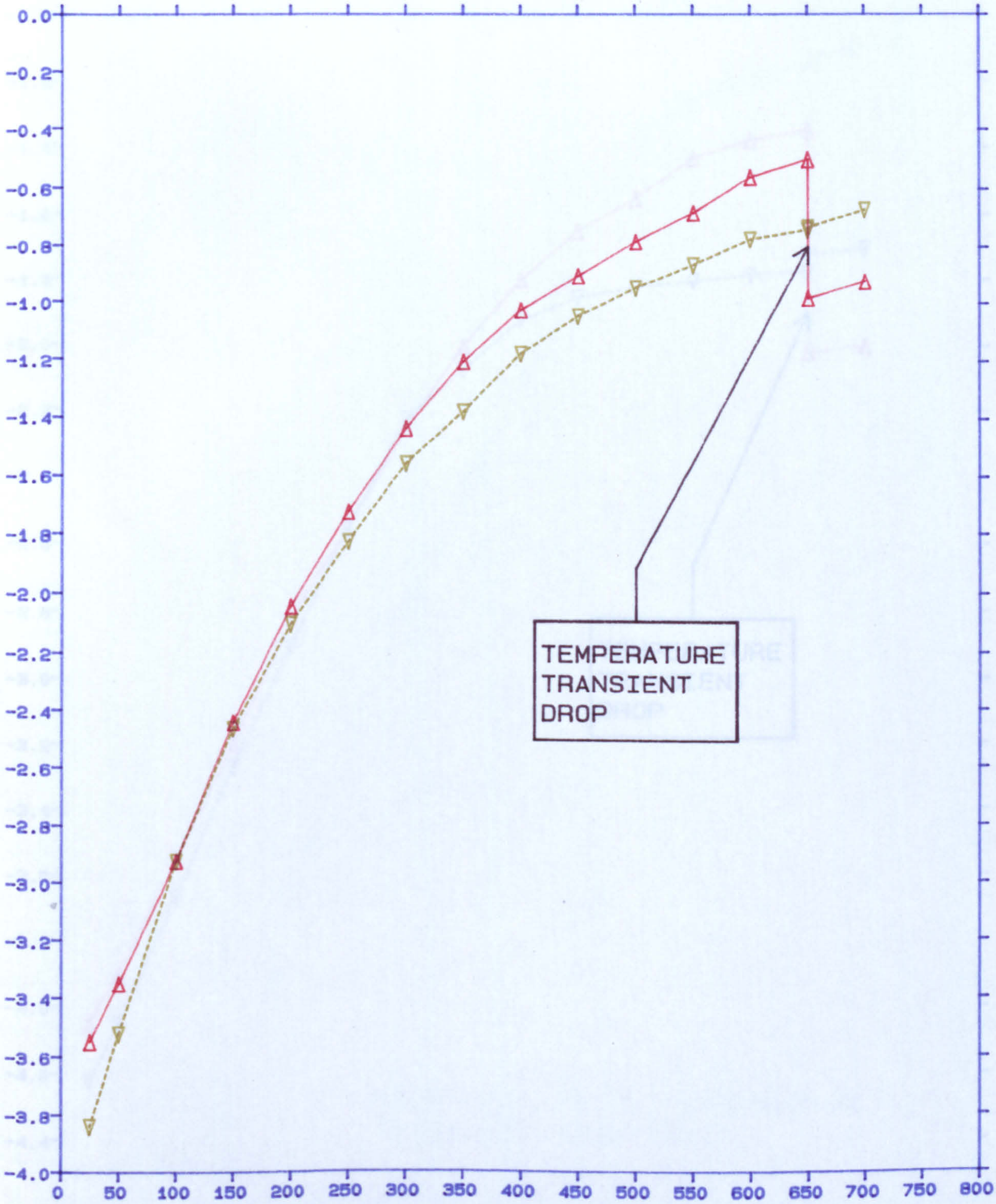


FIG. F8.38. GGRAN2TT INITIAL TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE

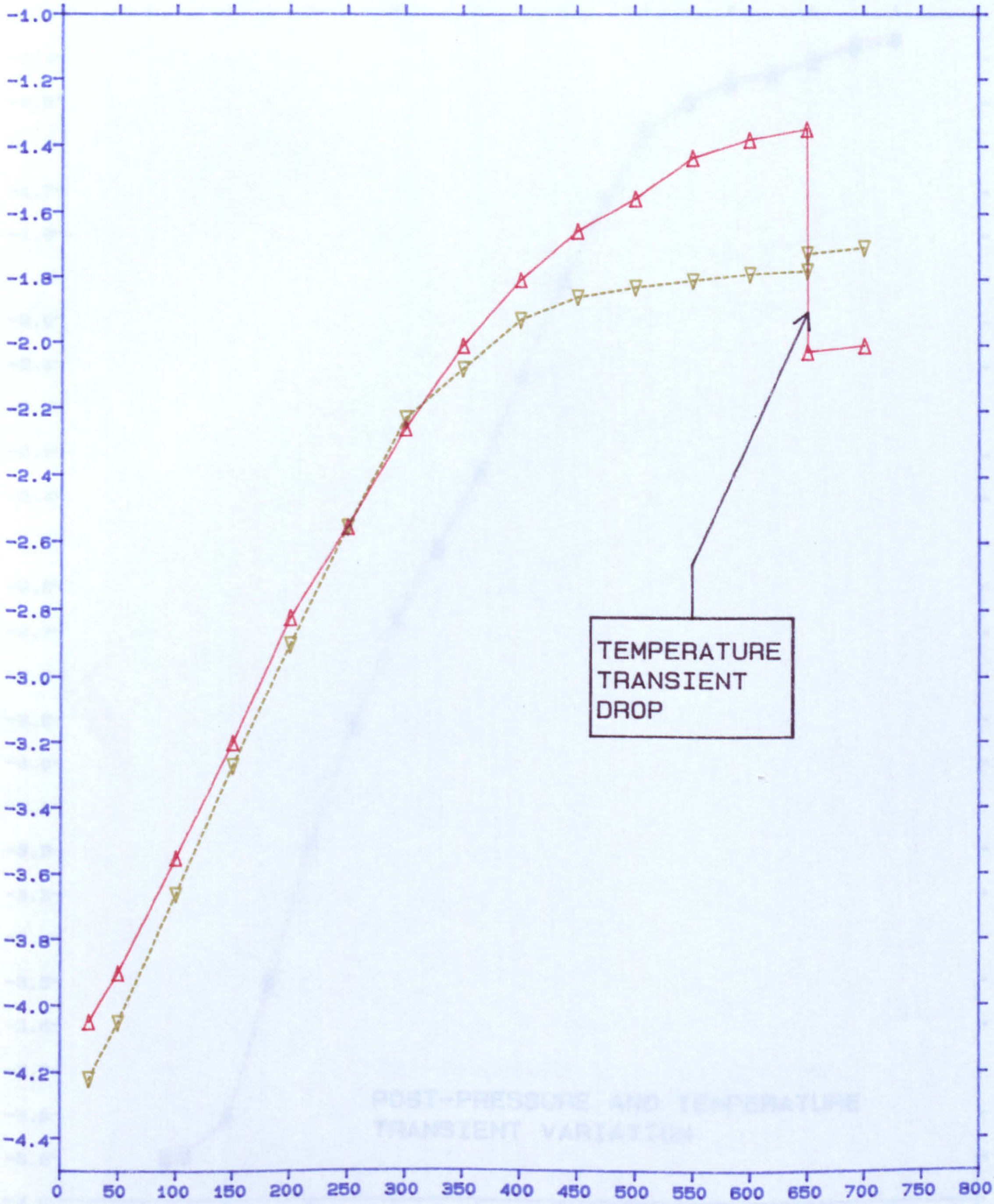


FIG. F8.39. IVD1T TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

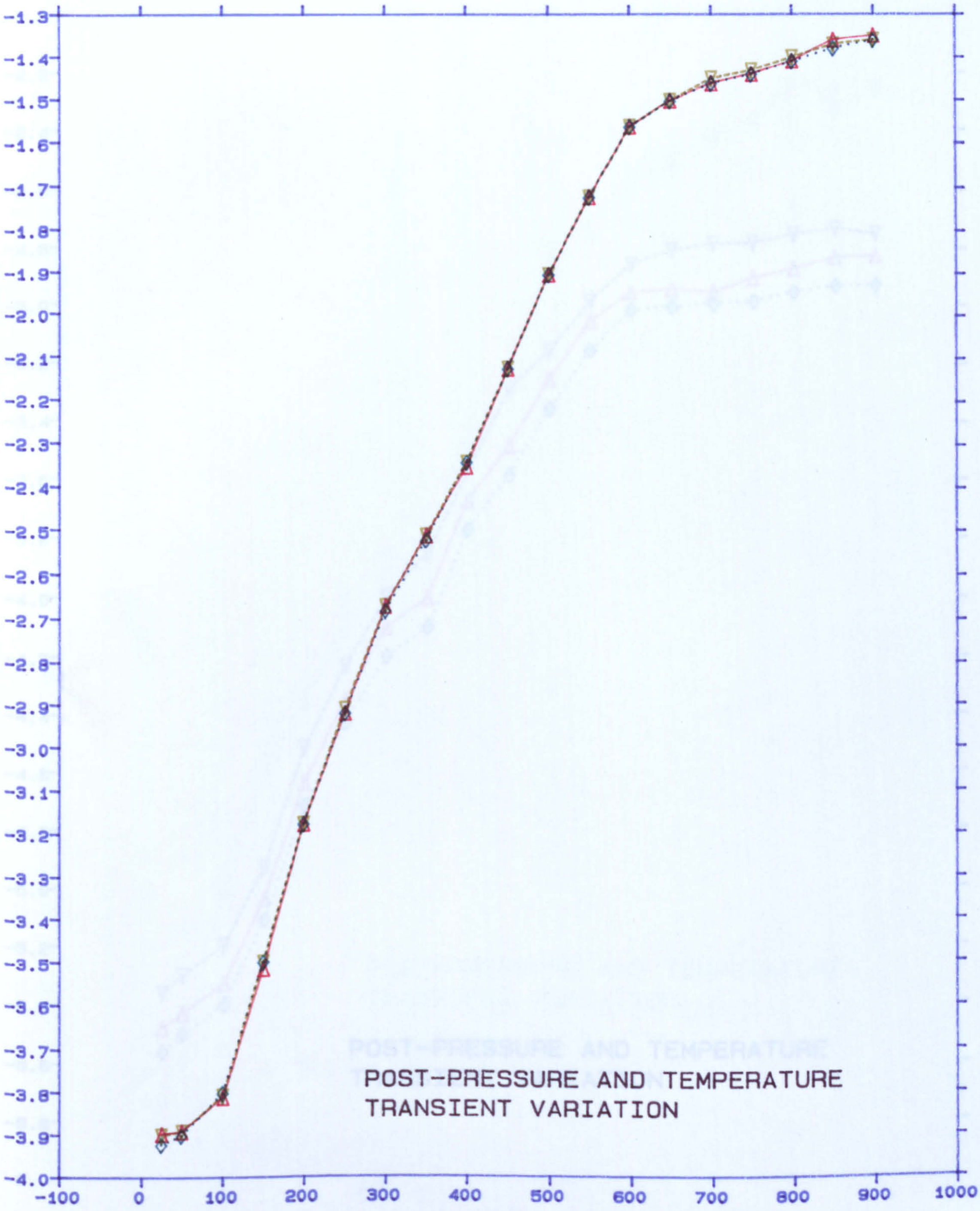


FIG. F8.40. IVD2T TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

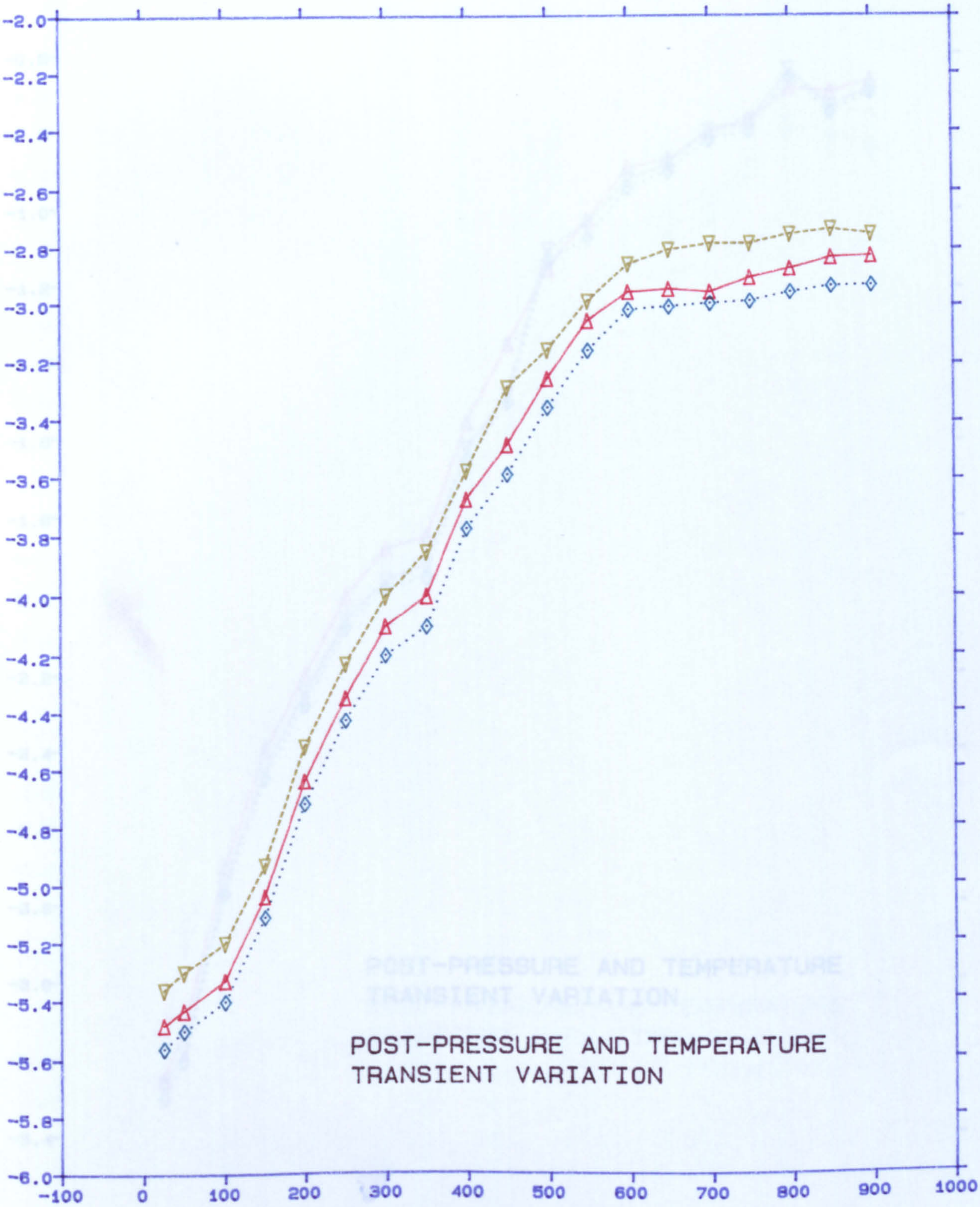


FIG. F8.41. IVE1T TEMPERATURE VARIATIONFOR 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

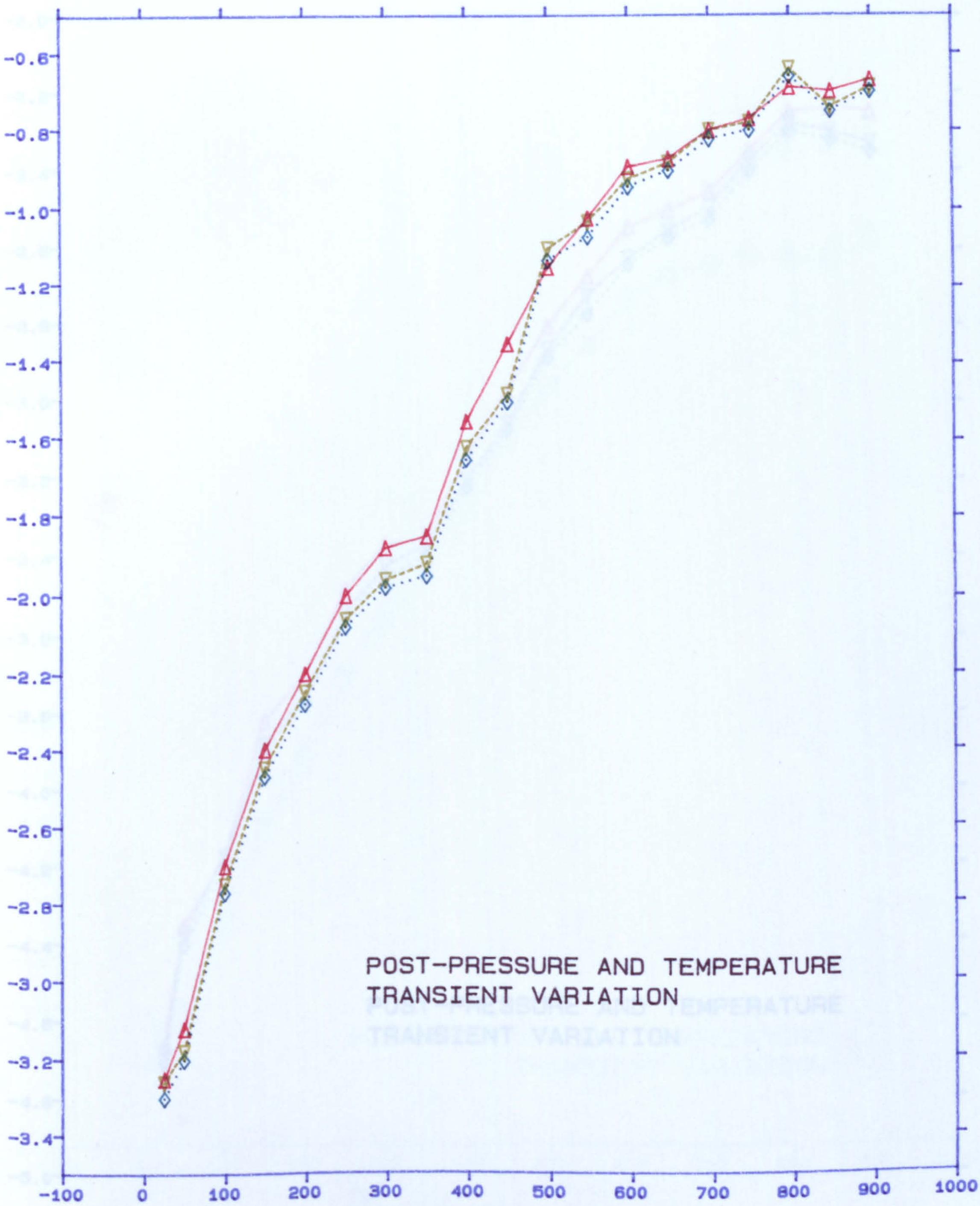


FIG. F8.42. IVE2T TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED BASIC GRANULITE

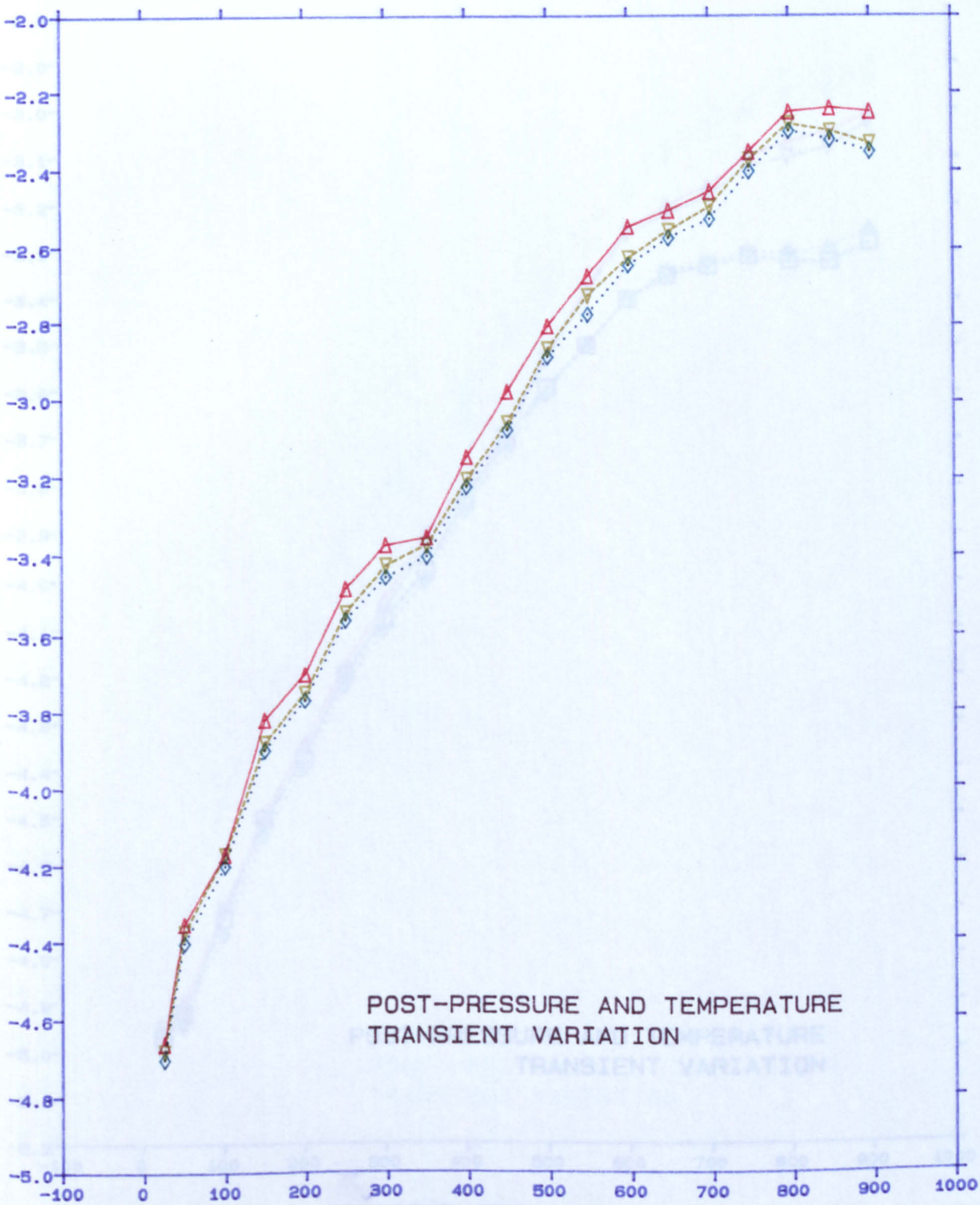


FIG. F8.43. MER21T TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE

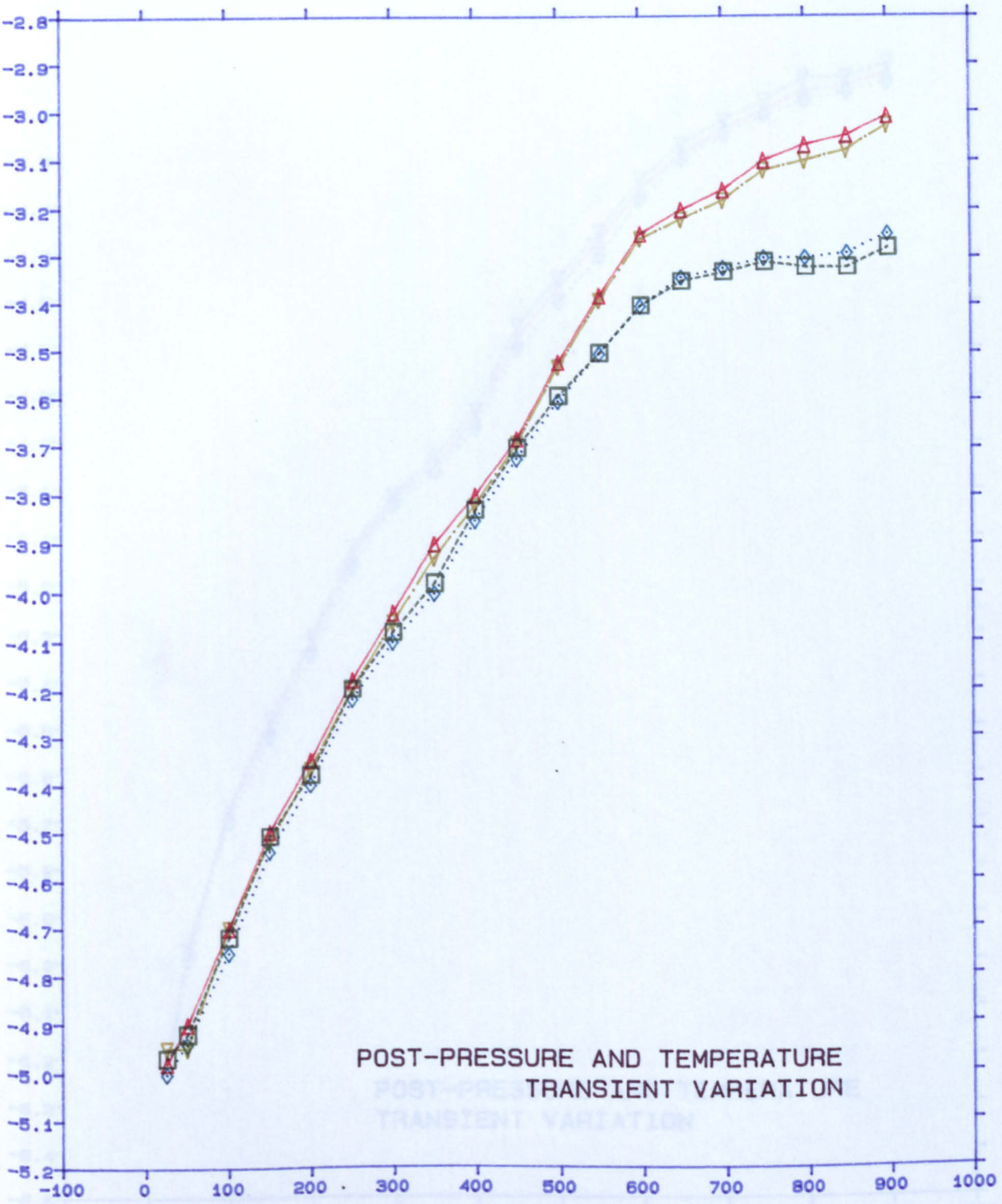


FIG. F8.44. MER22T TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED ORTHOPYROXENITE

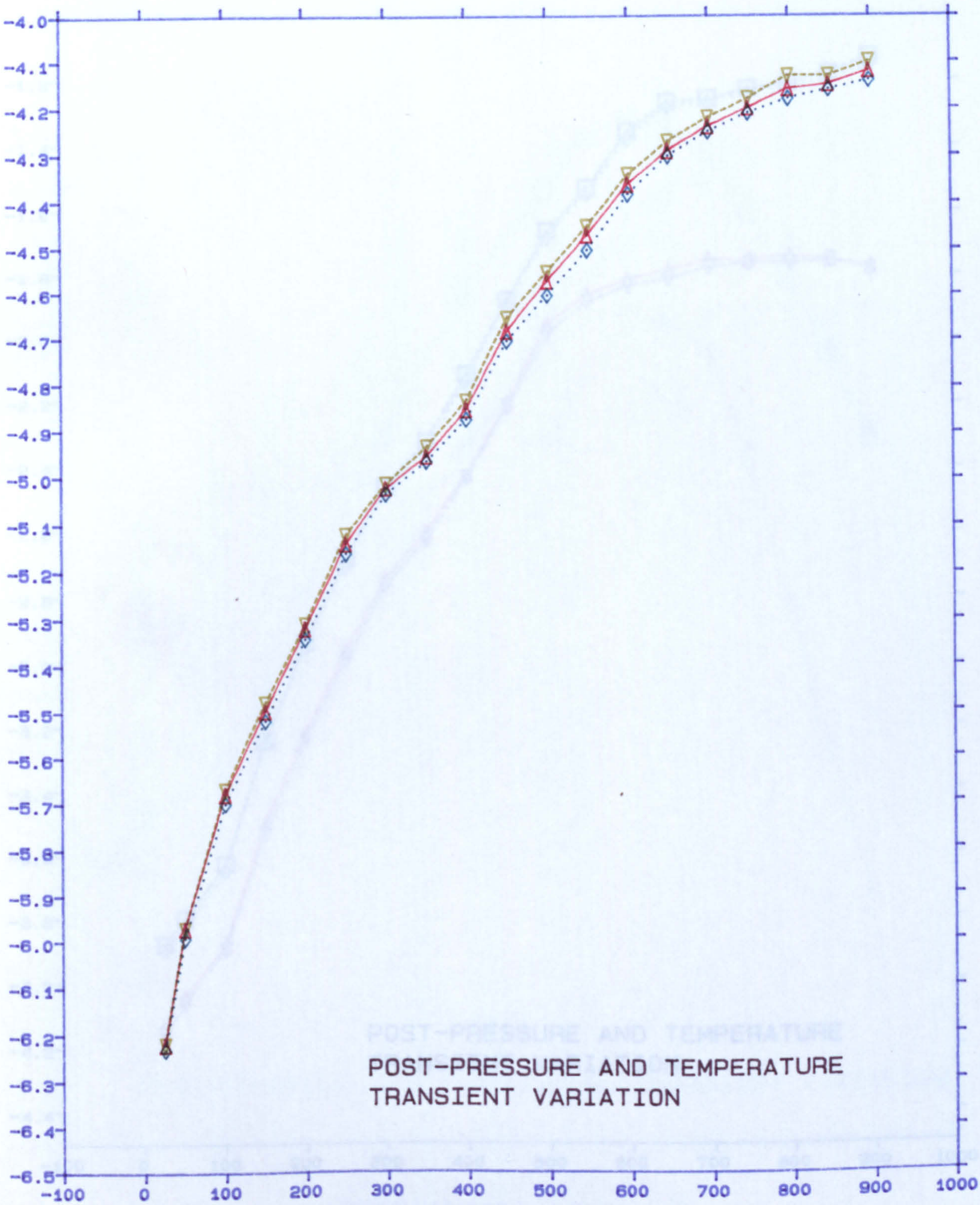


FIG. F8.45. A8BT TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

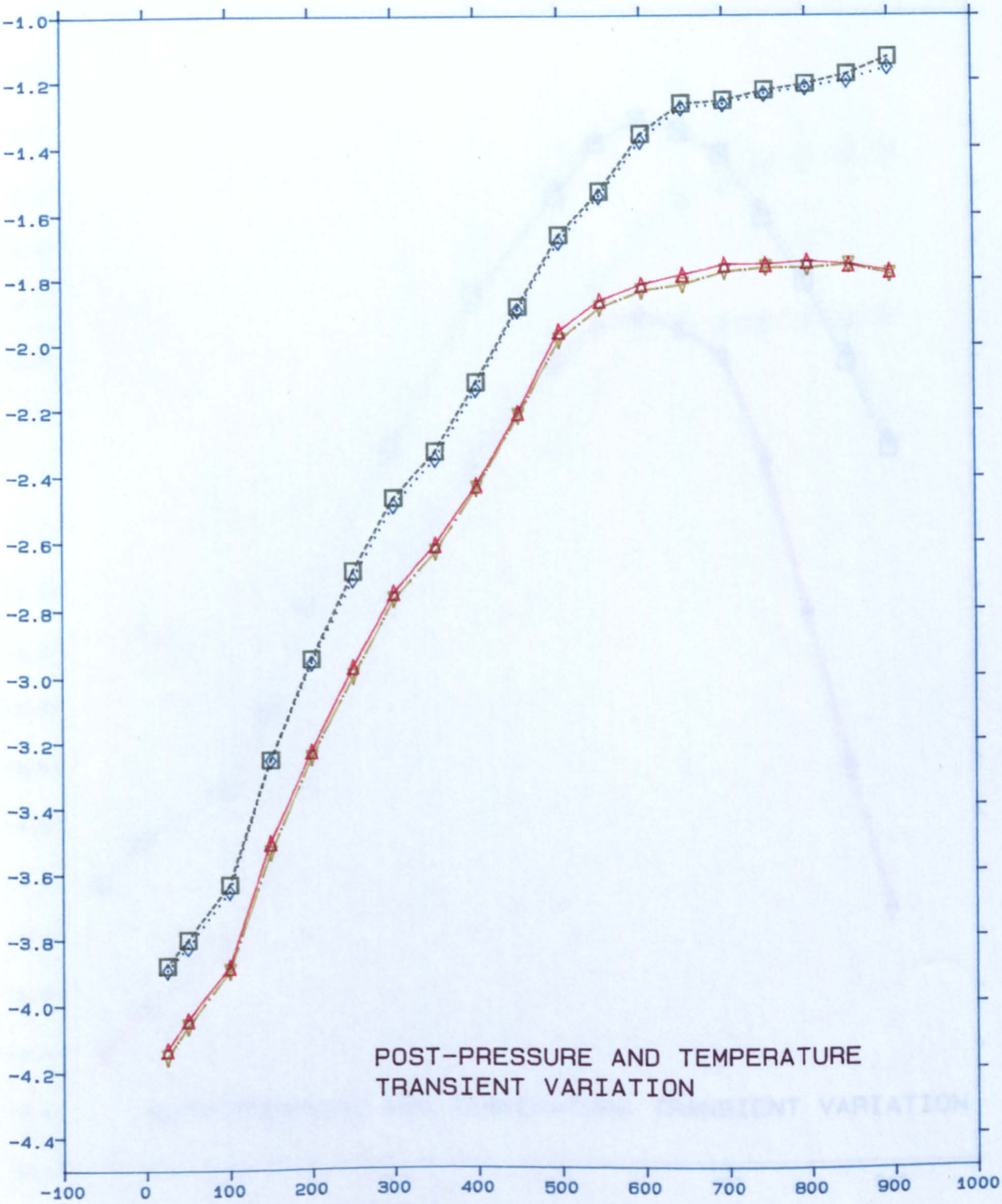


FIG. F8.46. L33BT TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE

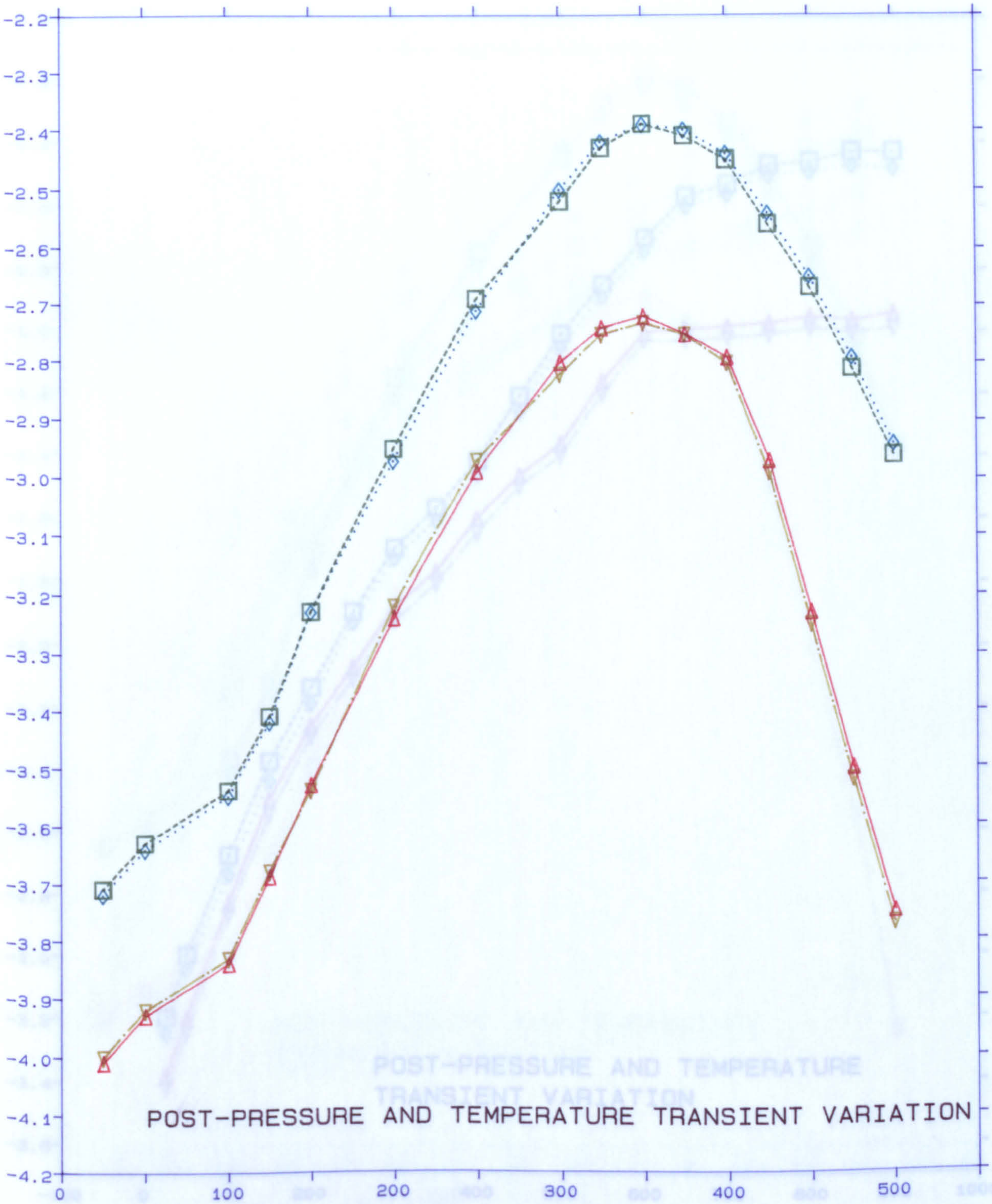


FIG. F8.47. L32T TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

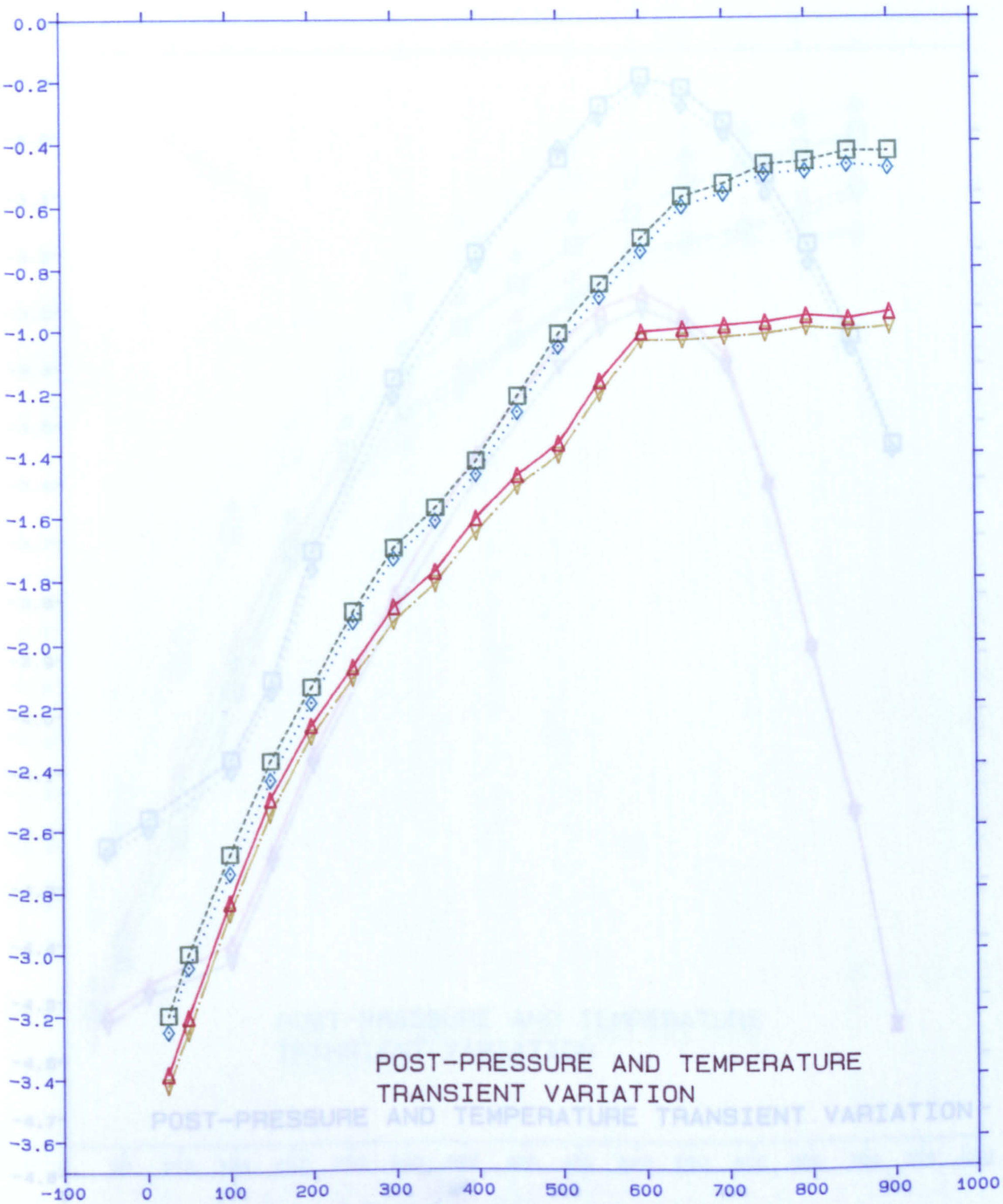


FIG. F8.48. V1T TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANITE

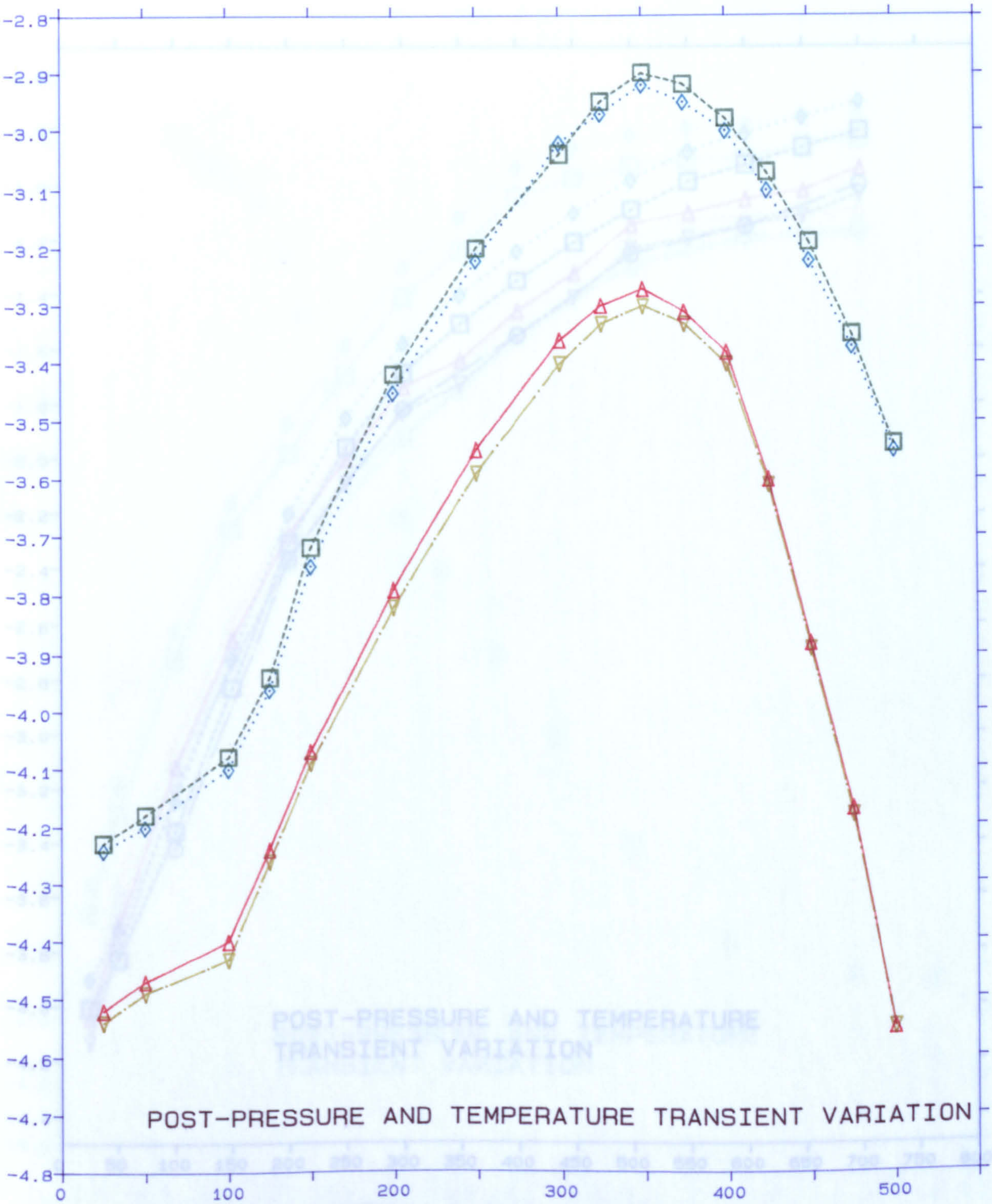


FIG. F8.49. GGRAN1T TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAHITE GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATIONS

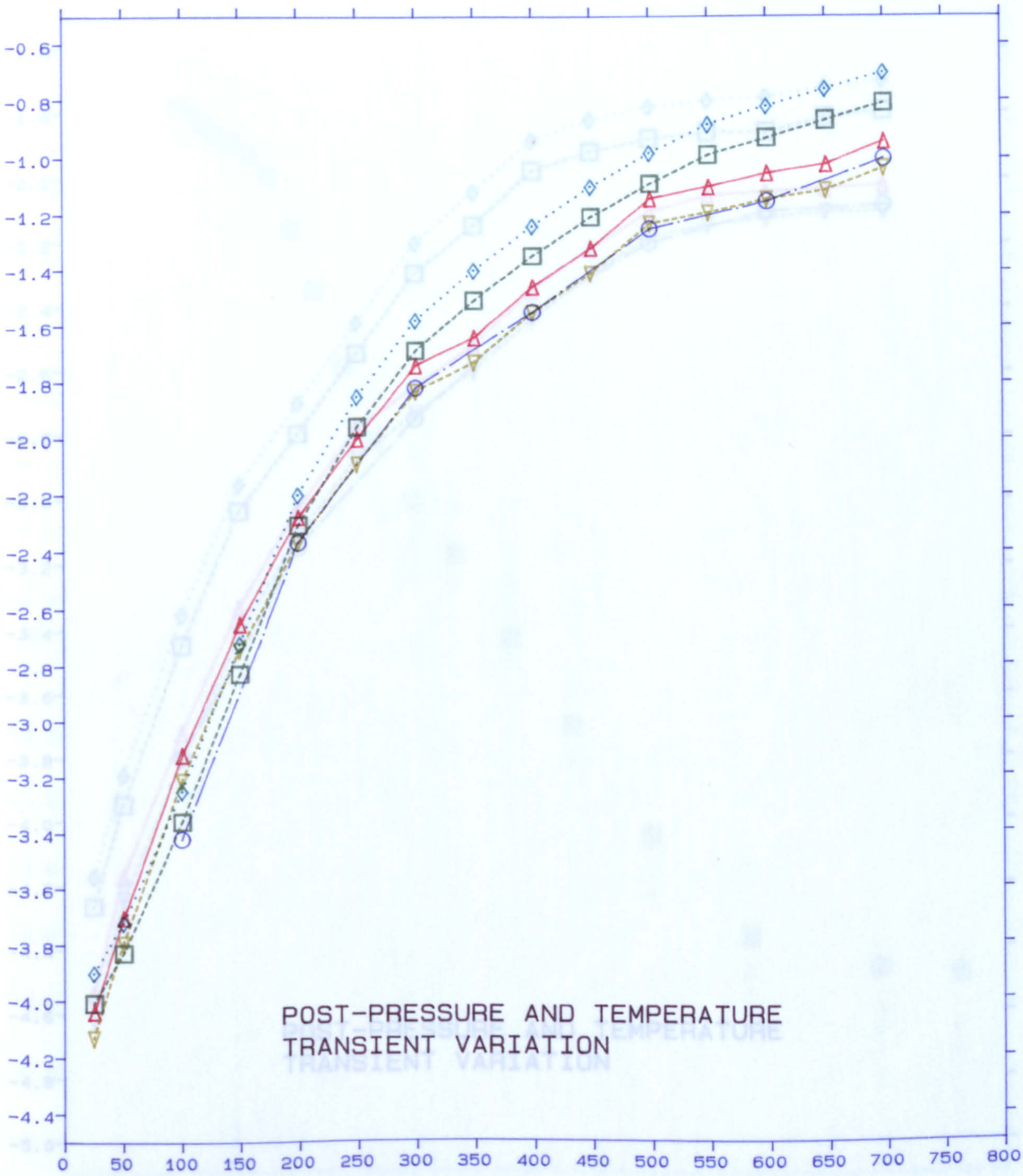


FIG. F8.50. GGRAN2T TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATION

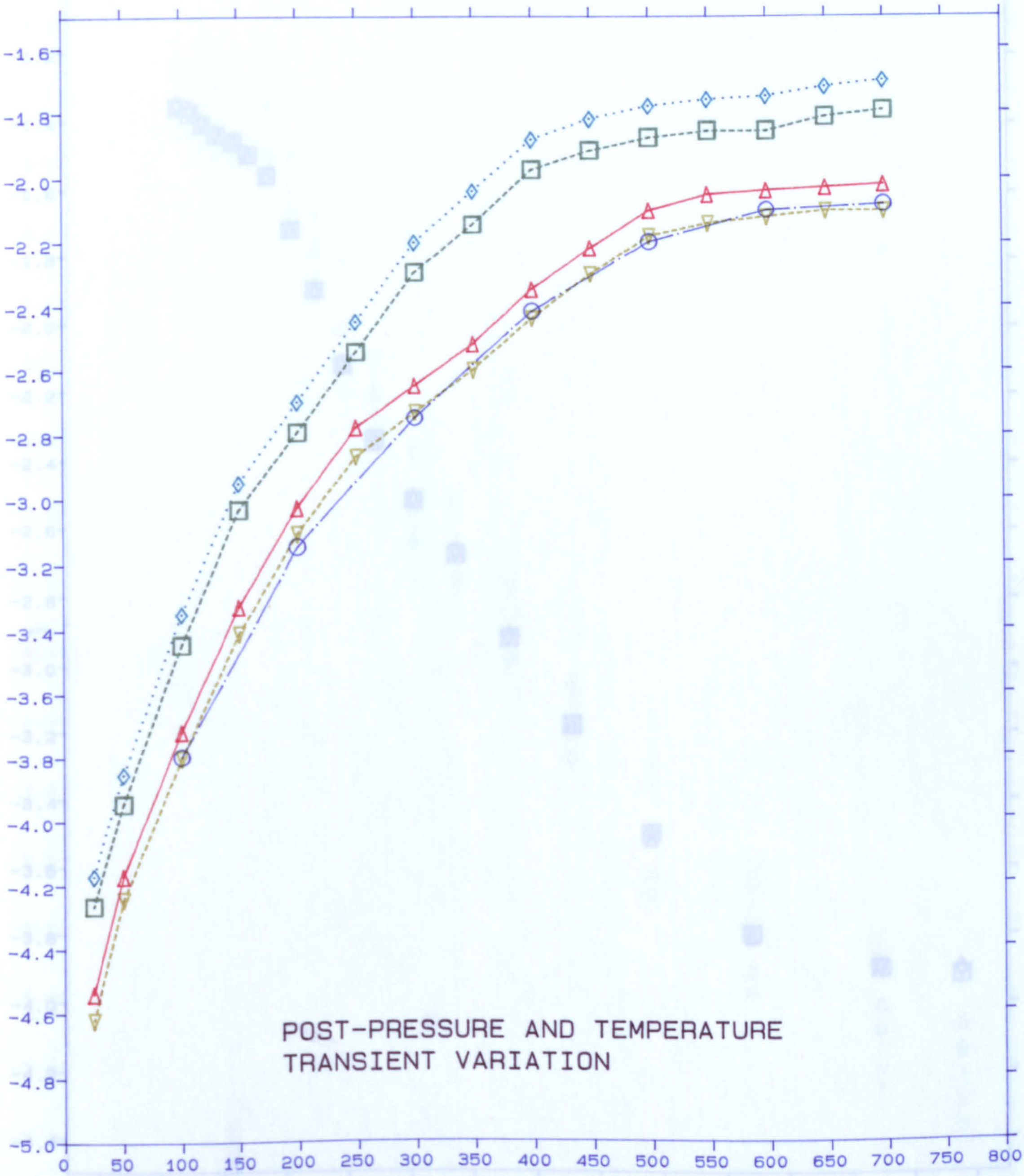


FIG. F8.51. IVD1TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

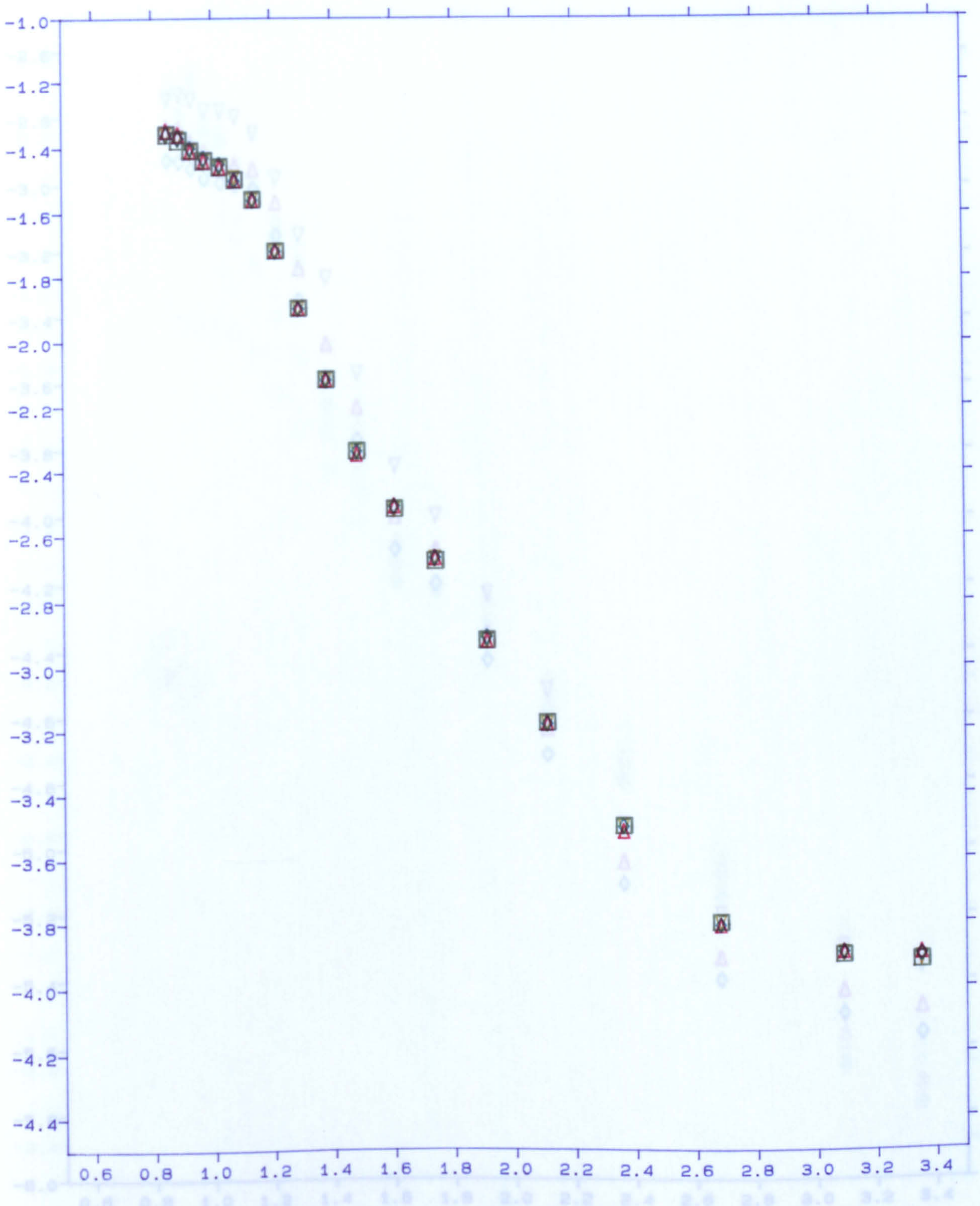


FIG. F8.52. IVD2TK INVERSE TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

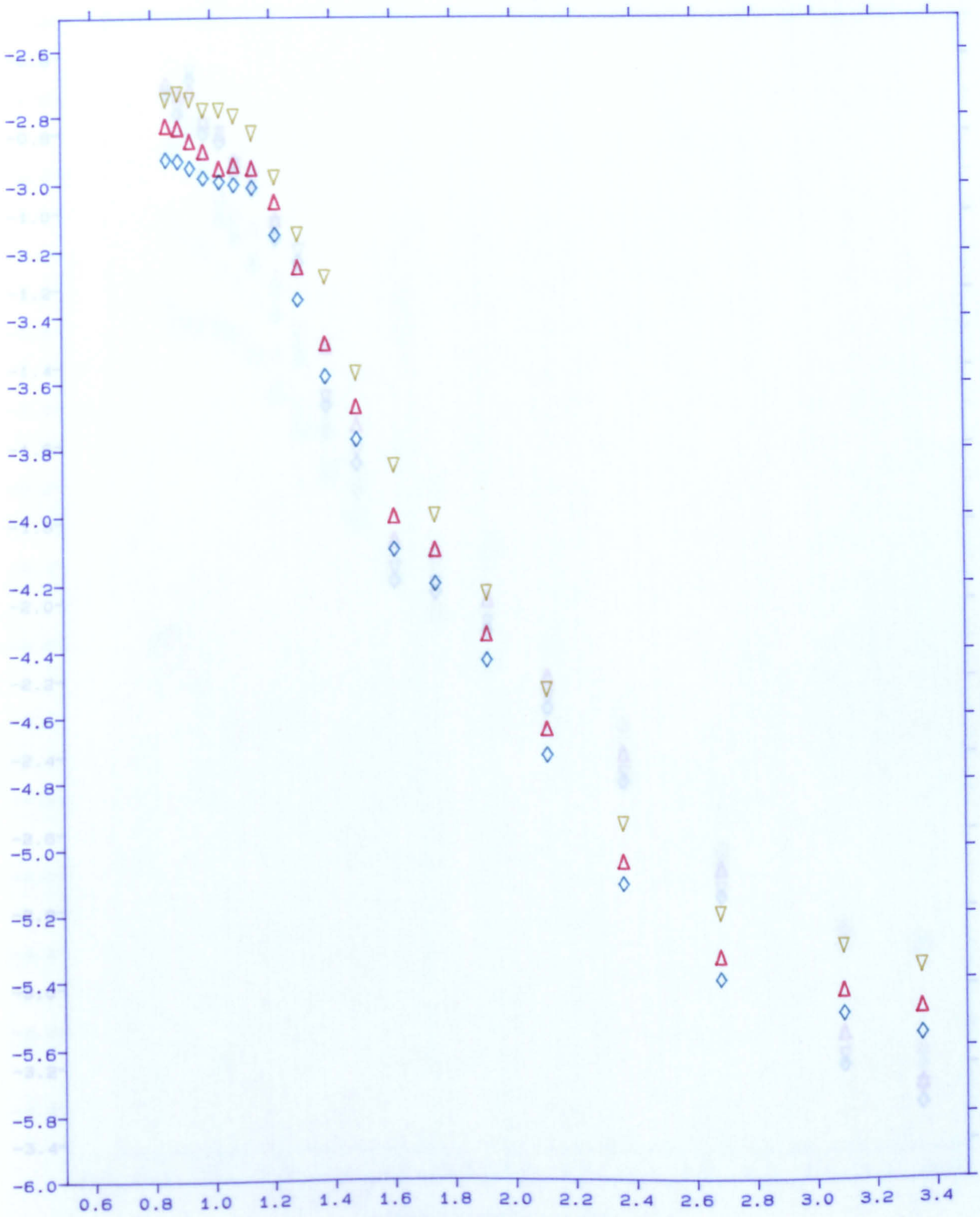


FIG. F8.53. IVE1TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

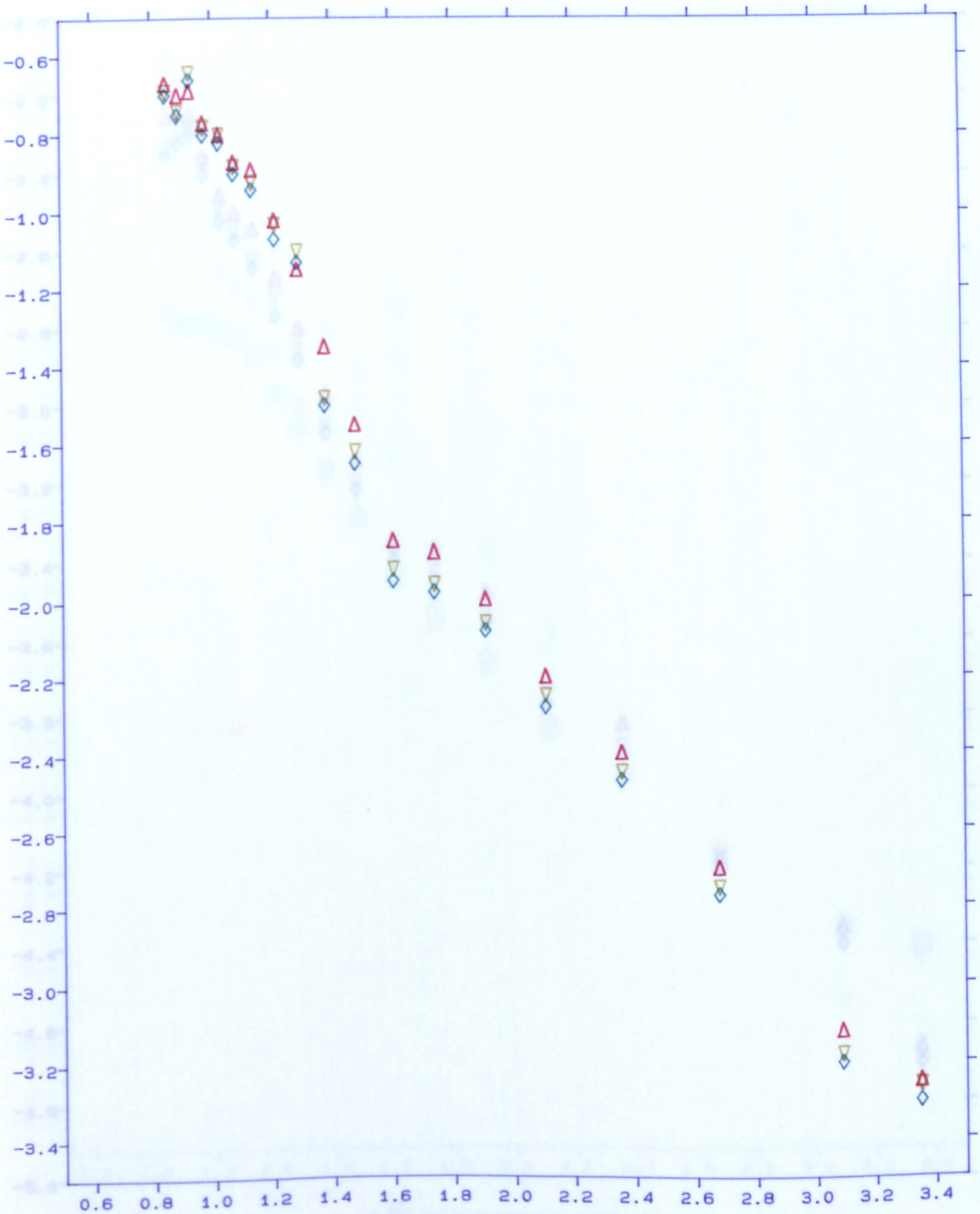


FIG. F8.54. IVE2TK INVERSE TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED BASIC GRANULITE

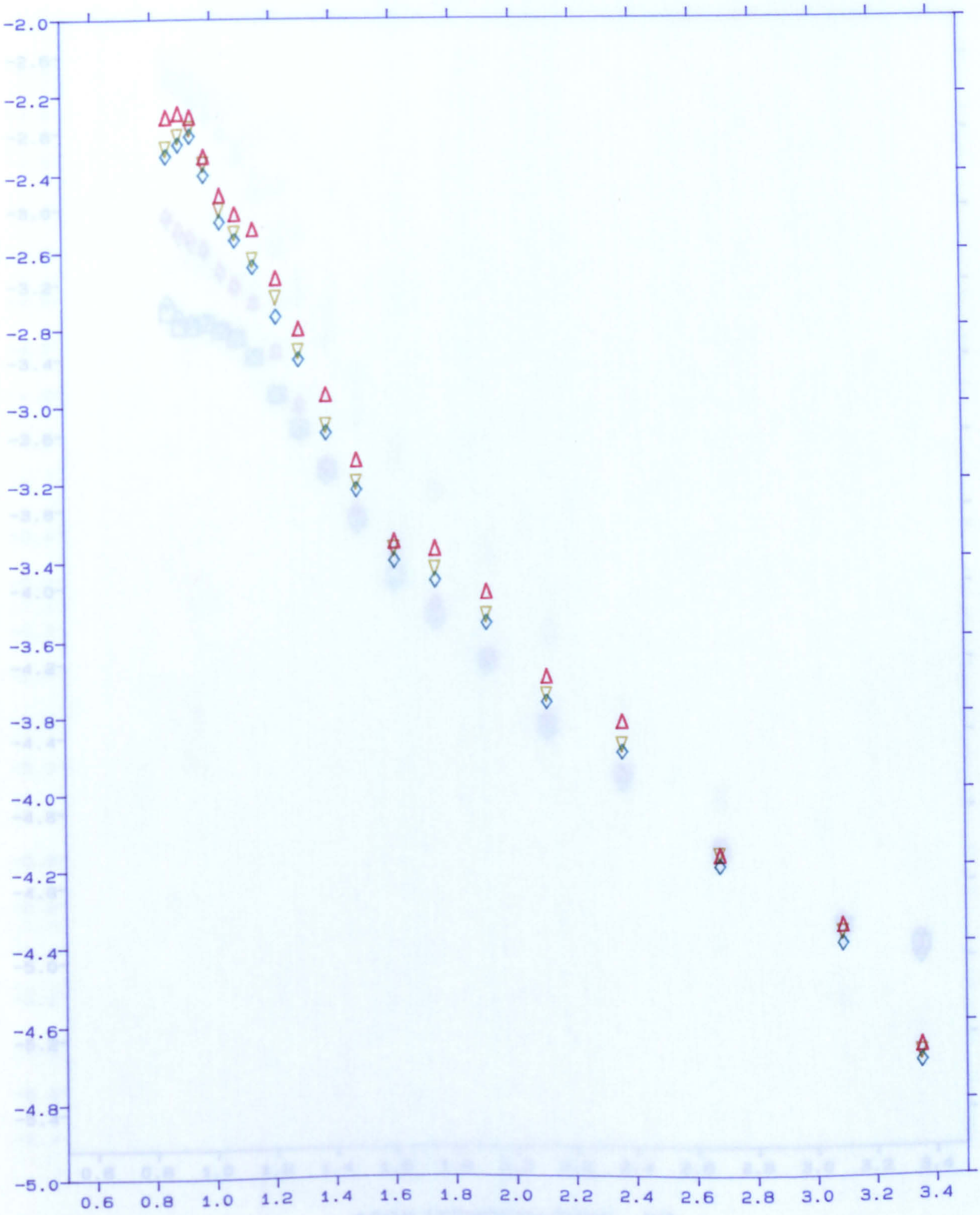


FIG. F8.55. MER21TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE

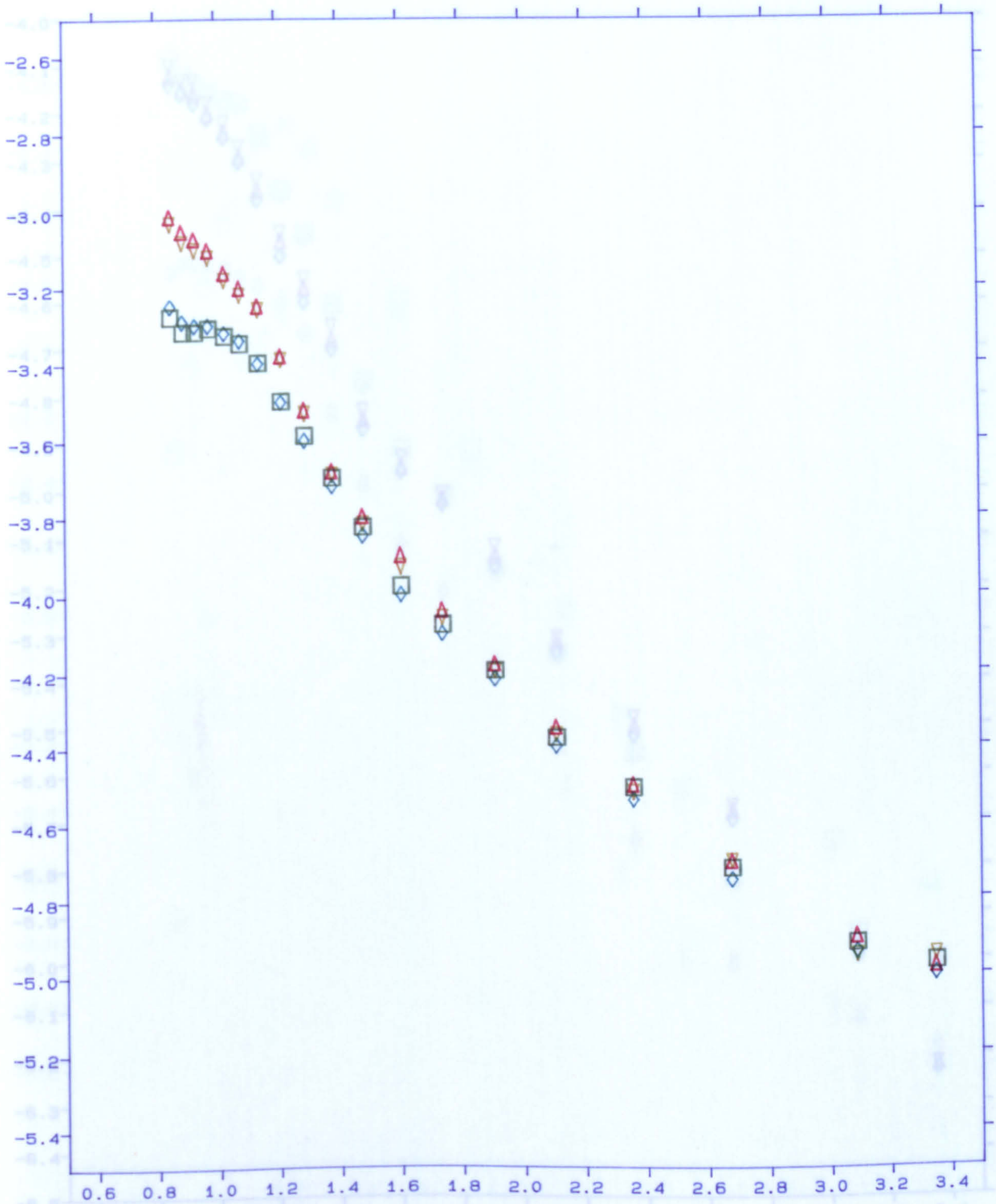


FIG. F8.56. MER22TK INVERSE TEMPERATURE VARIATION FOR DISTILLED WATER SATURATED ORTHOPYROXENITE

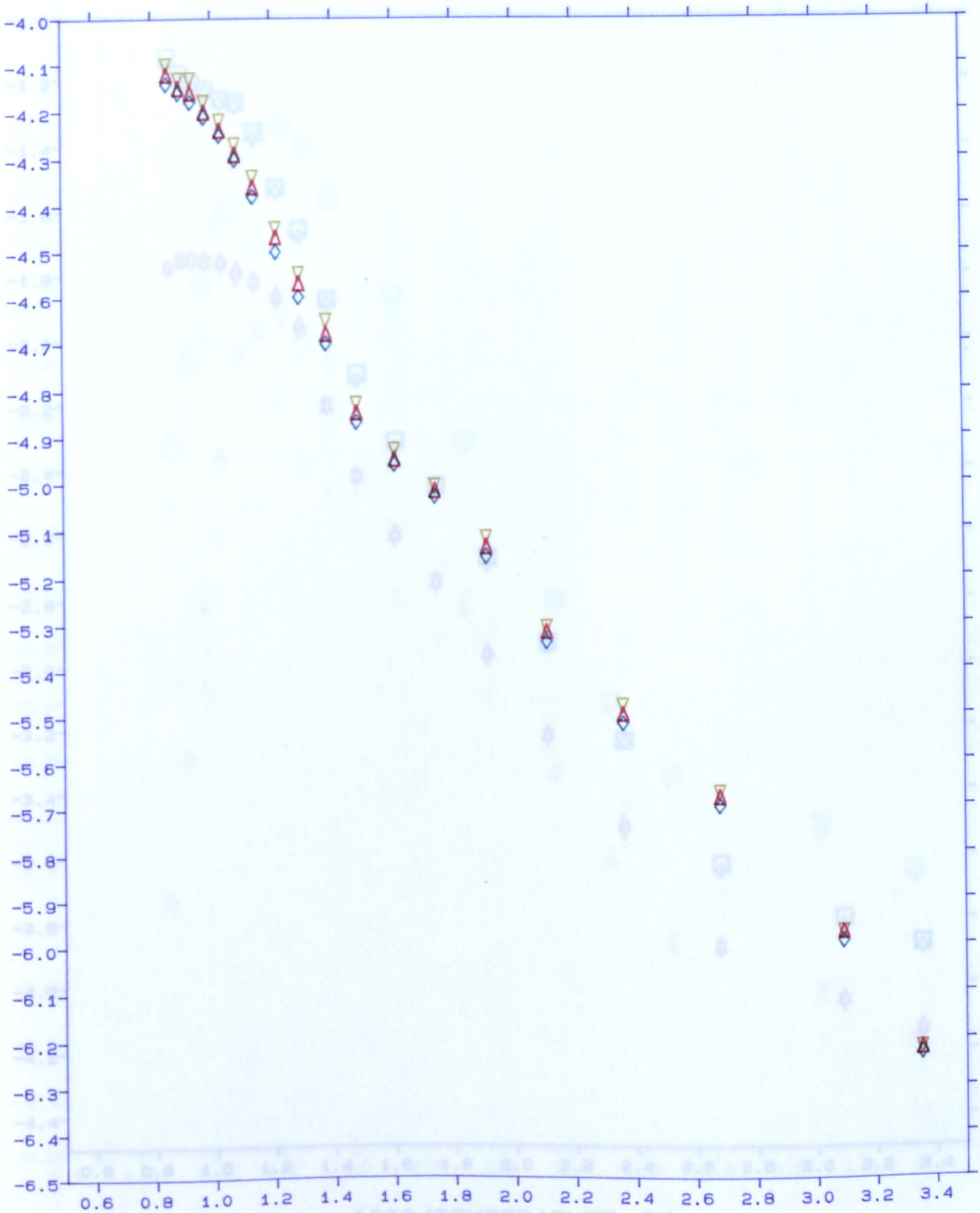


FIG. F8.57. A8BTK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

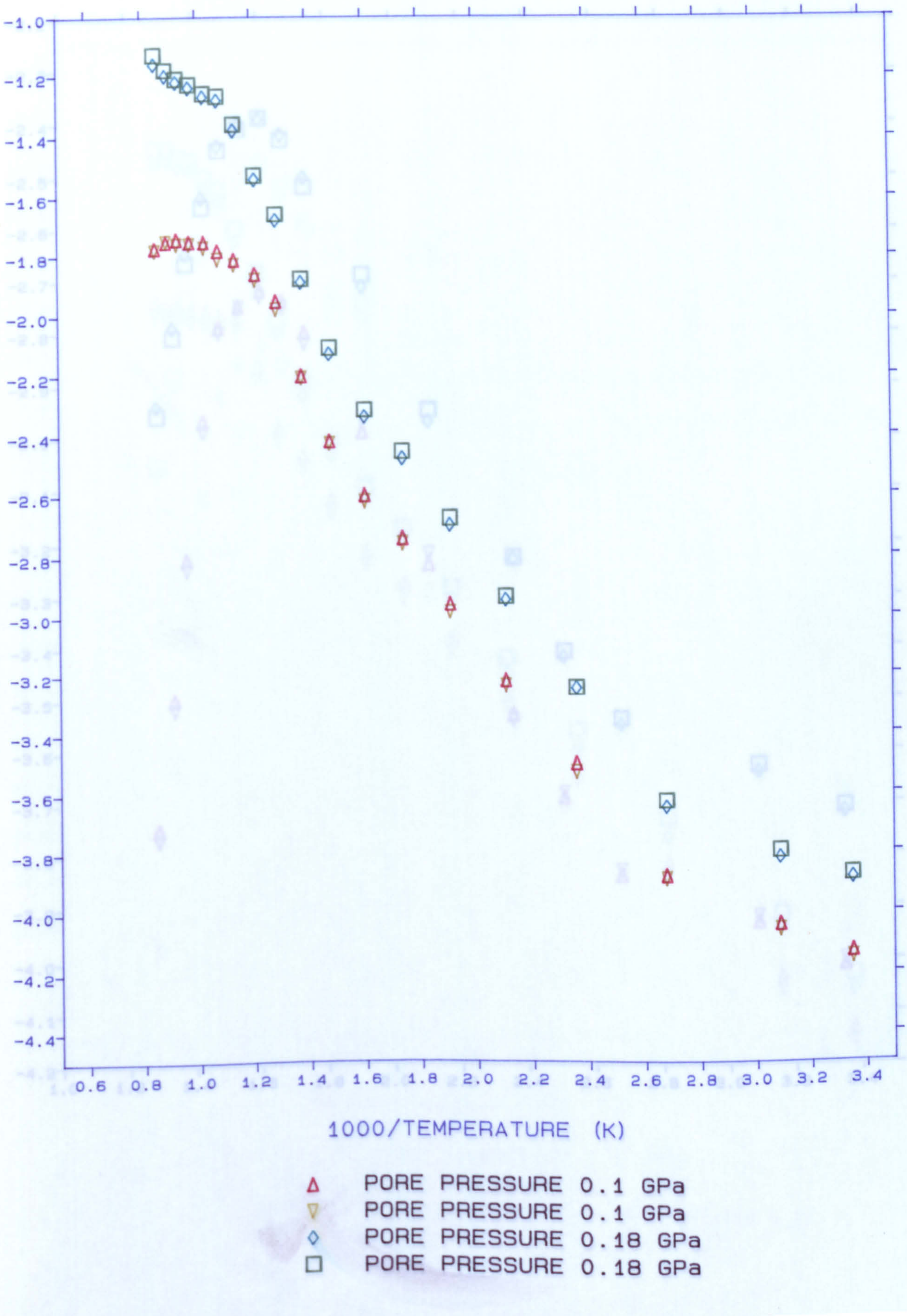


FIG. F8.58. L33BTK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE

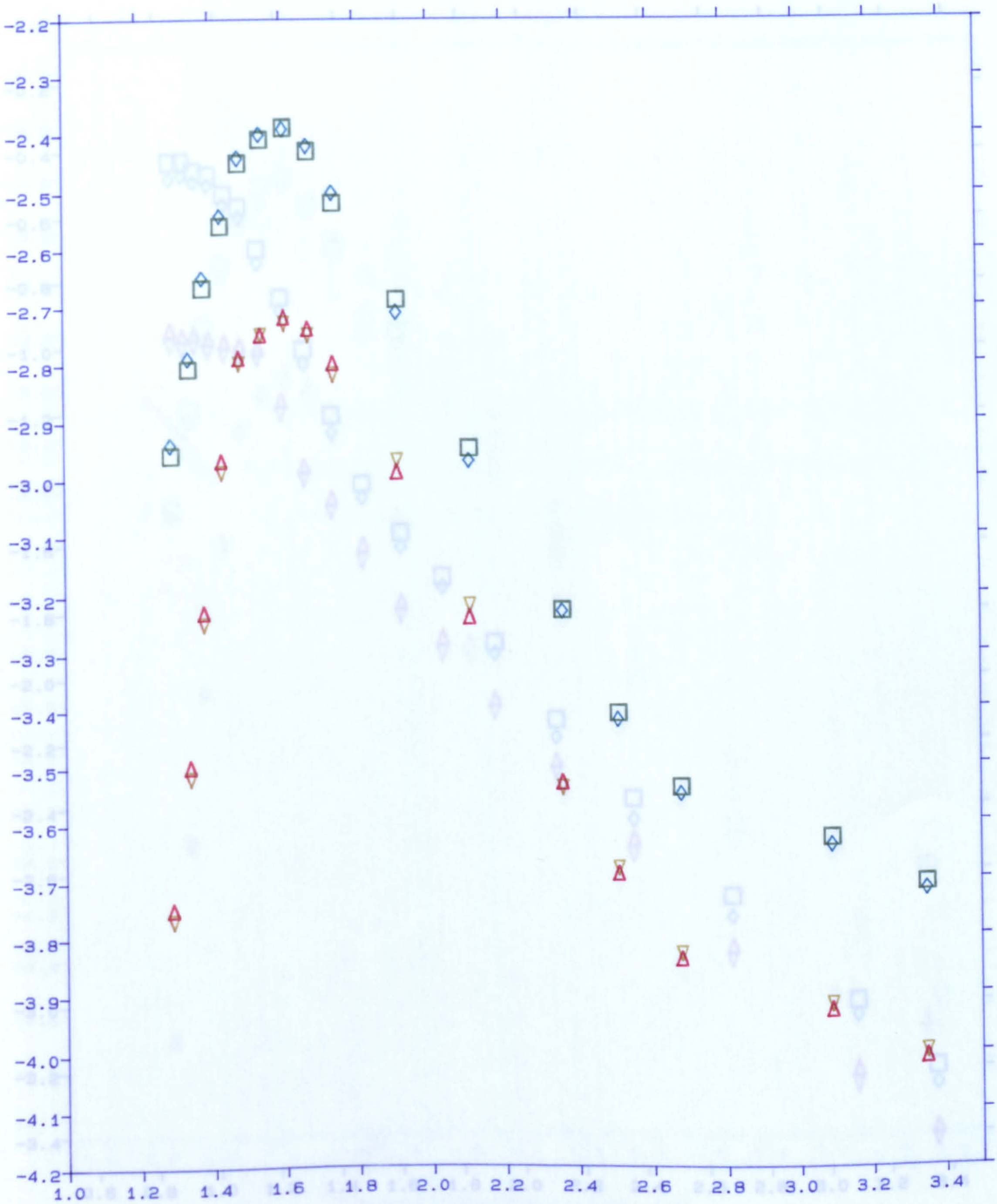


FIG. F8.59. L32TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

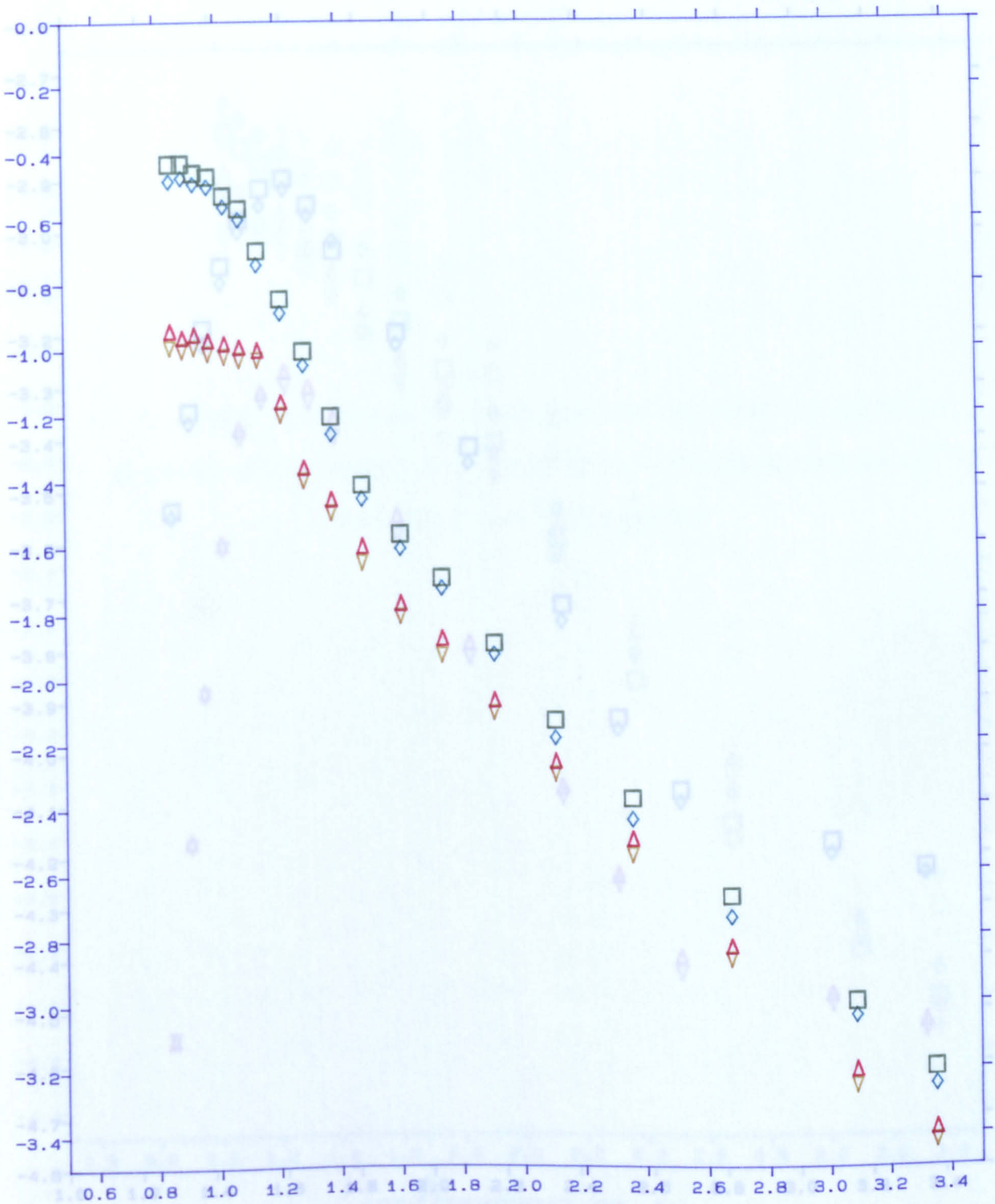


FIG. F8.60. V1TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANITE

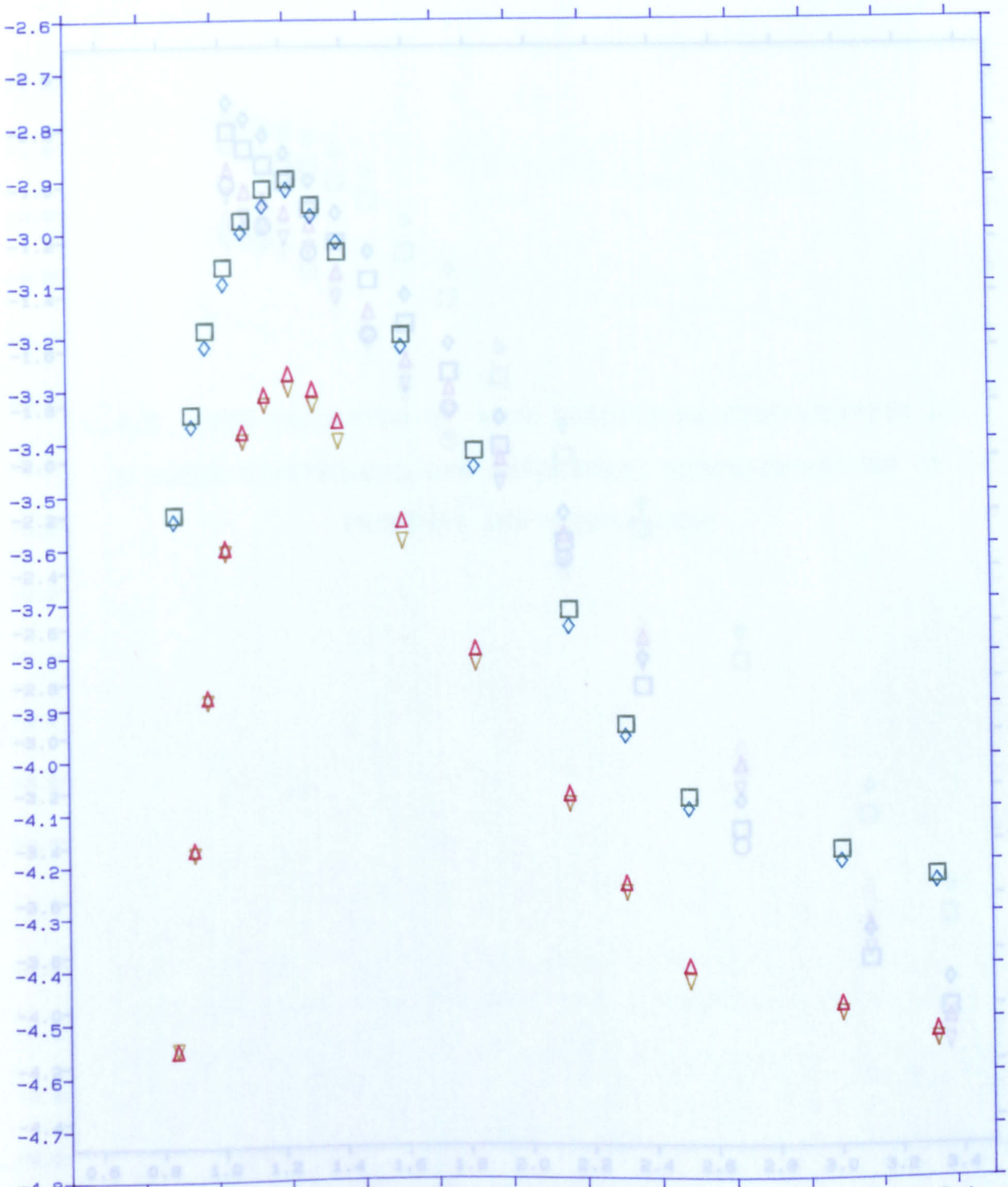
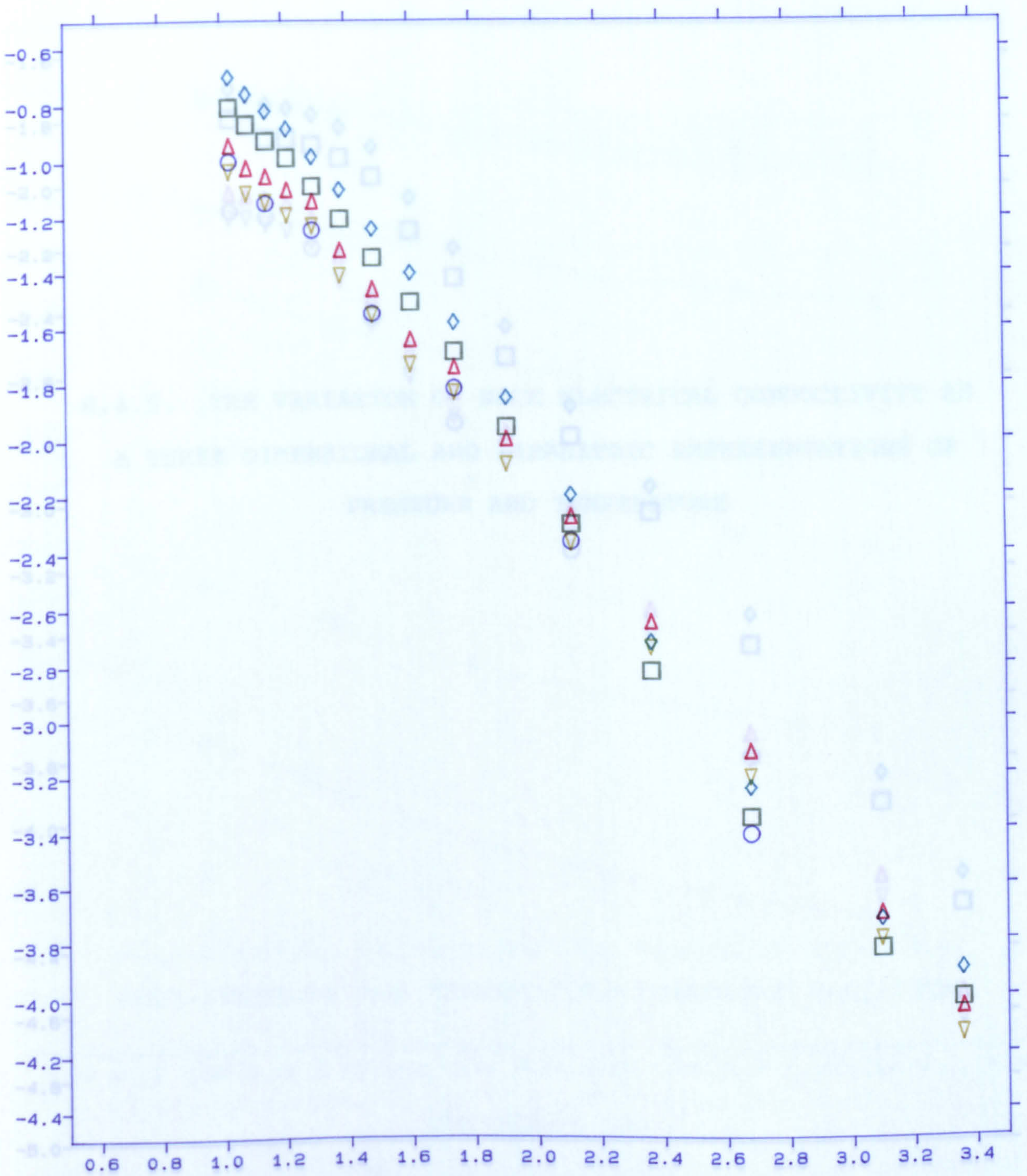
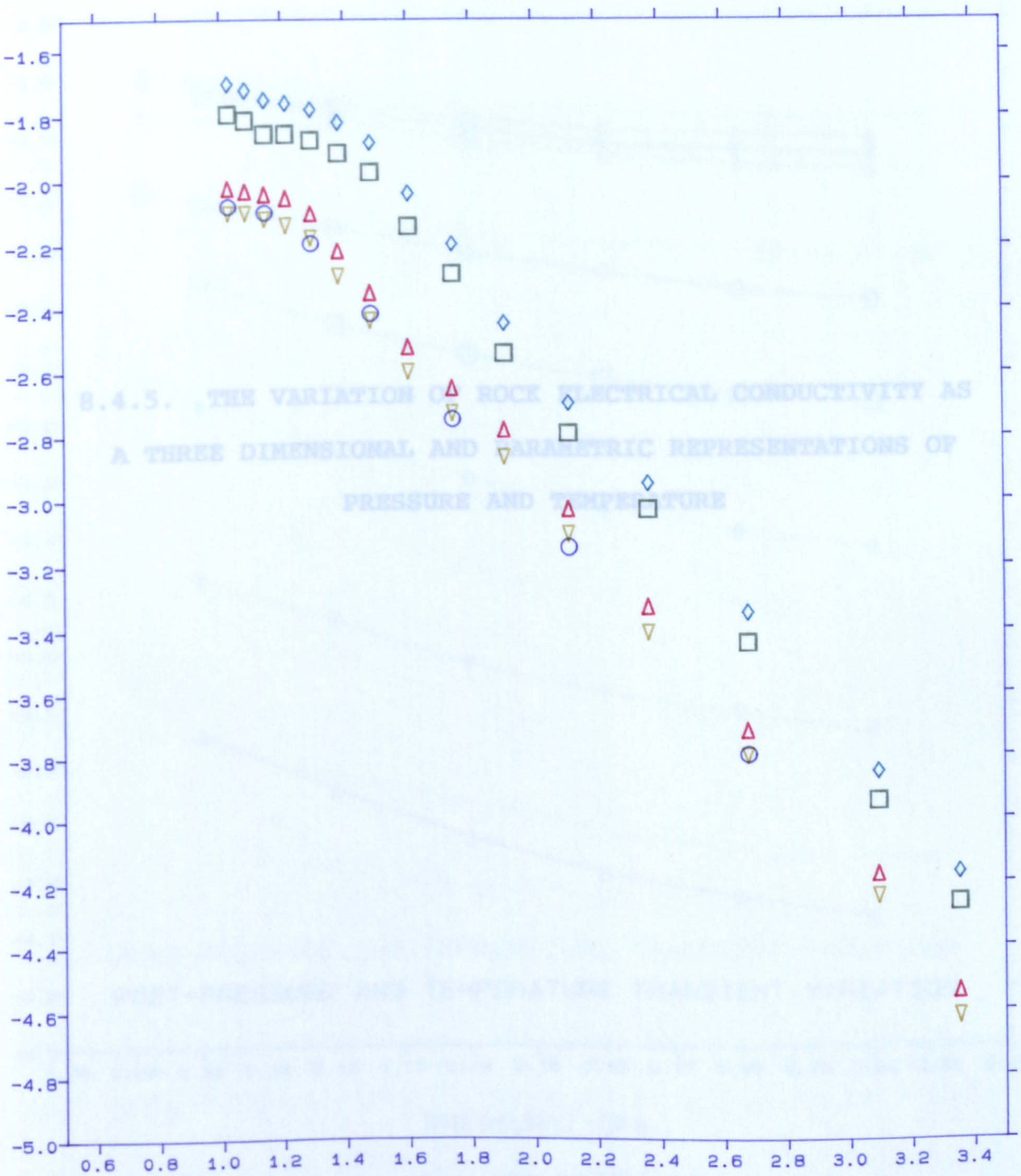


FIG. F8.61. GGRAN1TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATION



△ PORE PRESSURE 0.1 GPa
▽ PORE PRESSURE 0.1 GPa
◇ PORE PRESSURE 0.18 GPa
□ PORE PRESSURE 0.18 GPa
○ PORE PRESSURE 0.08 GPa

FIG. F8.62. GGRAN2TK INVERSE TEMPERATURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATION



8.4.5. THE VARIATION OF ROCK ELECTRICAL CONDUCTIVITY AS A THREE DIMENSIONAL AND PARAMETRIC REPRESENTATIONS OF PRESSURE AND TEMPERATURE

FIG. F8.63. MER21PT PRESSURE VARIATION OF 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE

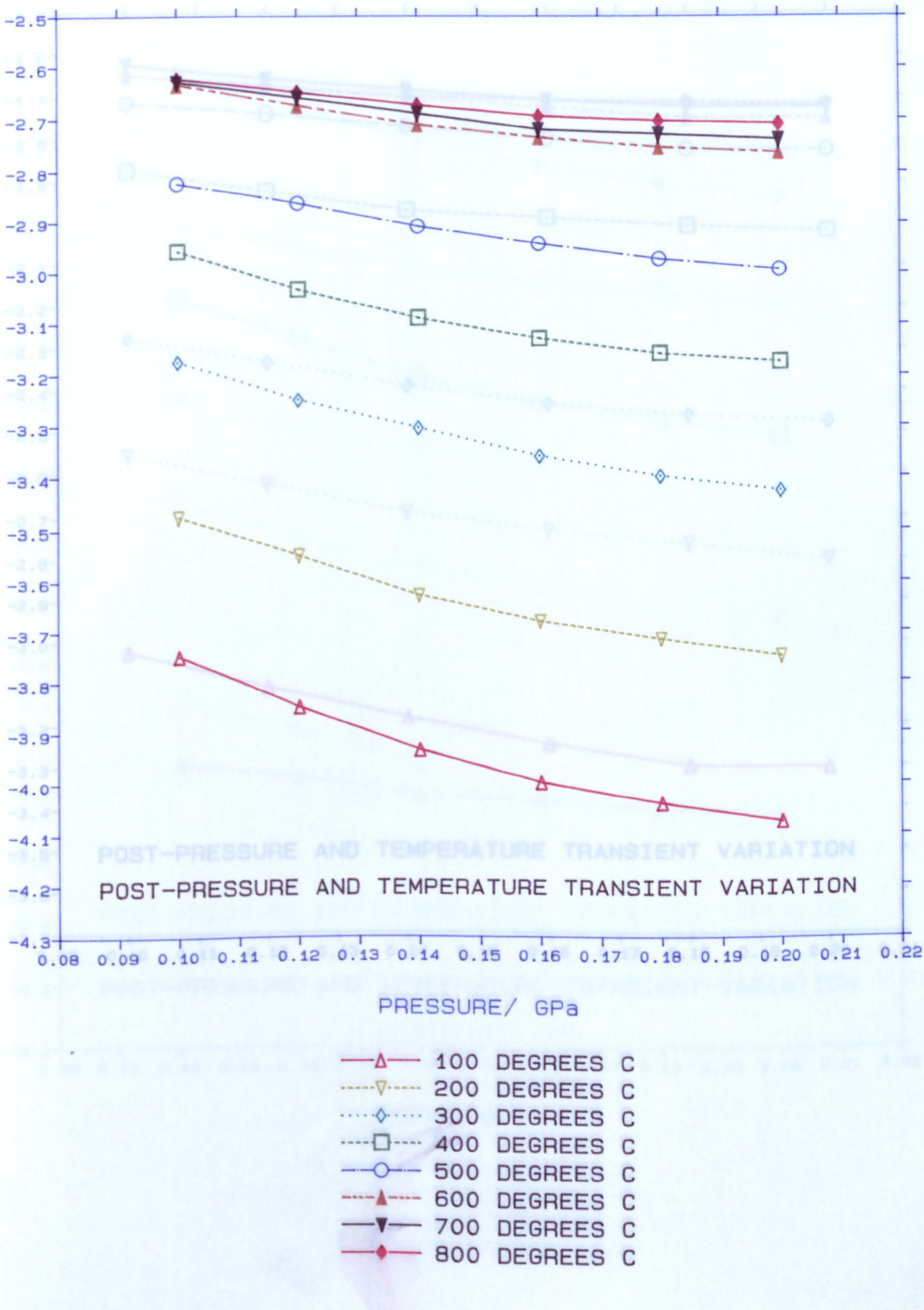


FIG. F8.64. A8BPT PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE

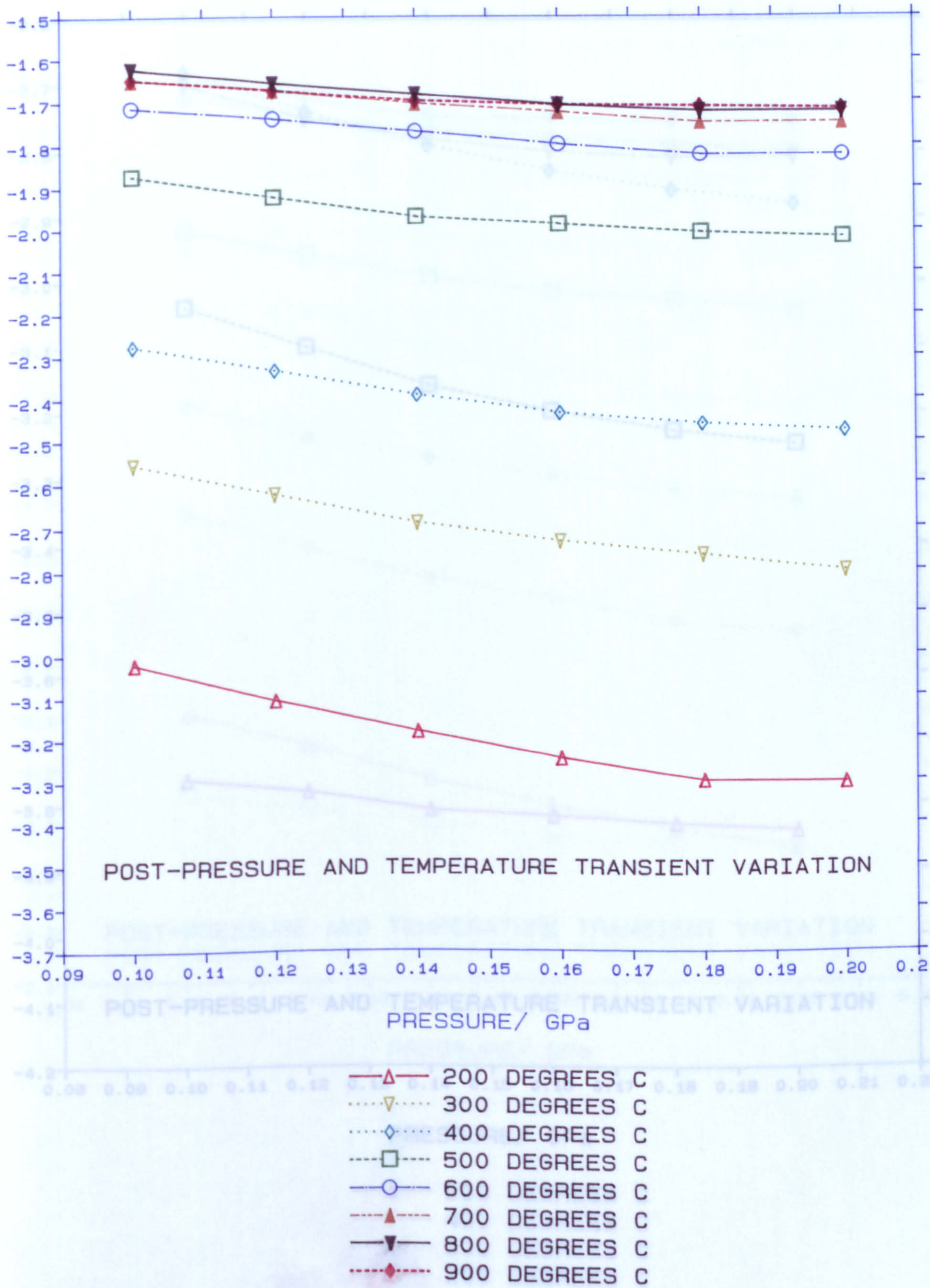


FIG. F8.65. L33BPT PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE

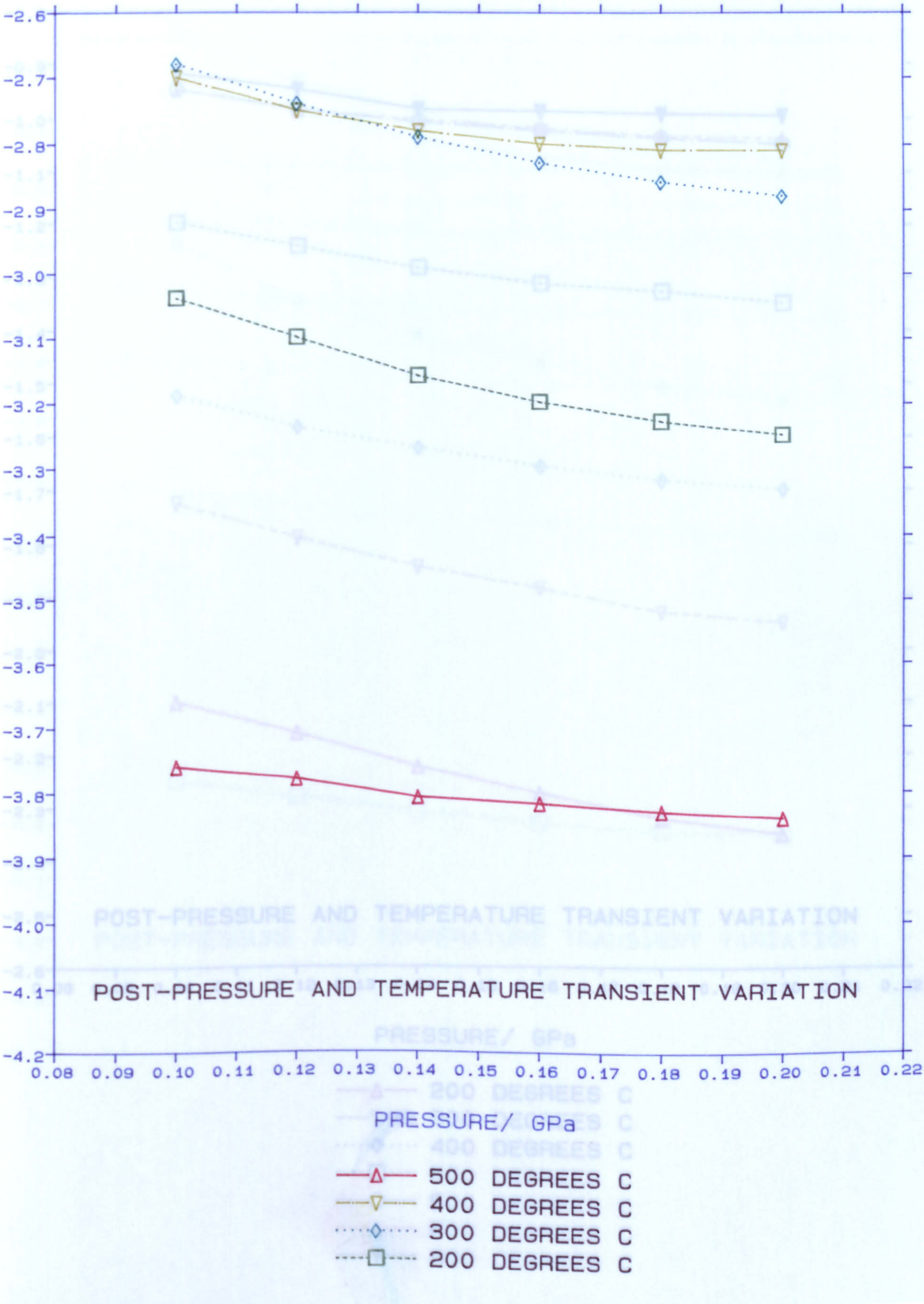


FIG. F8.66. L32PT PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE

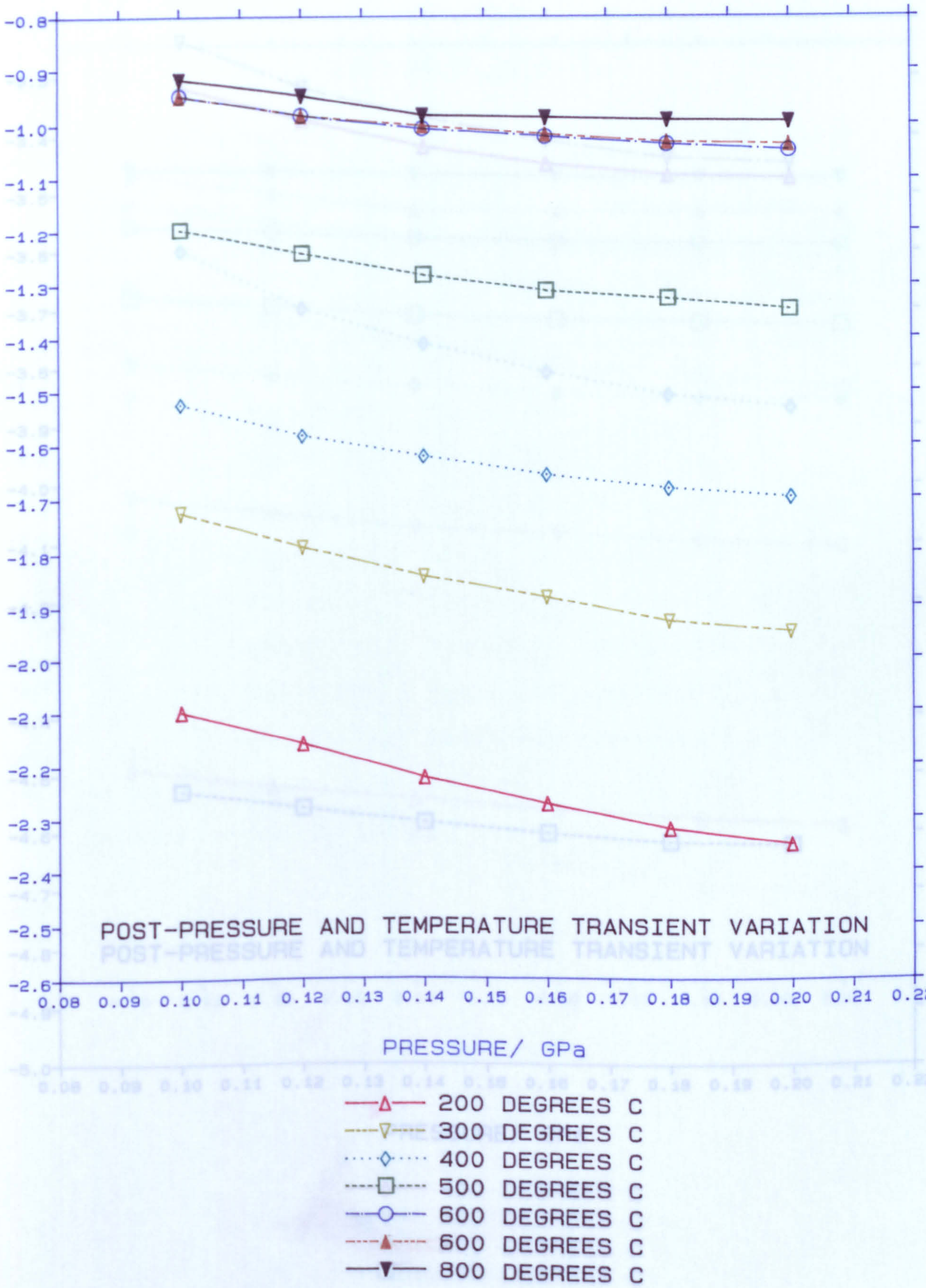


FIG. F8.67. V1PT PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANITE AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE

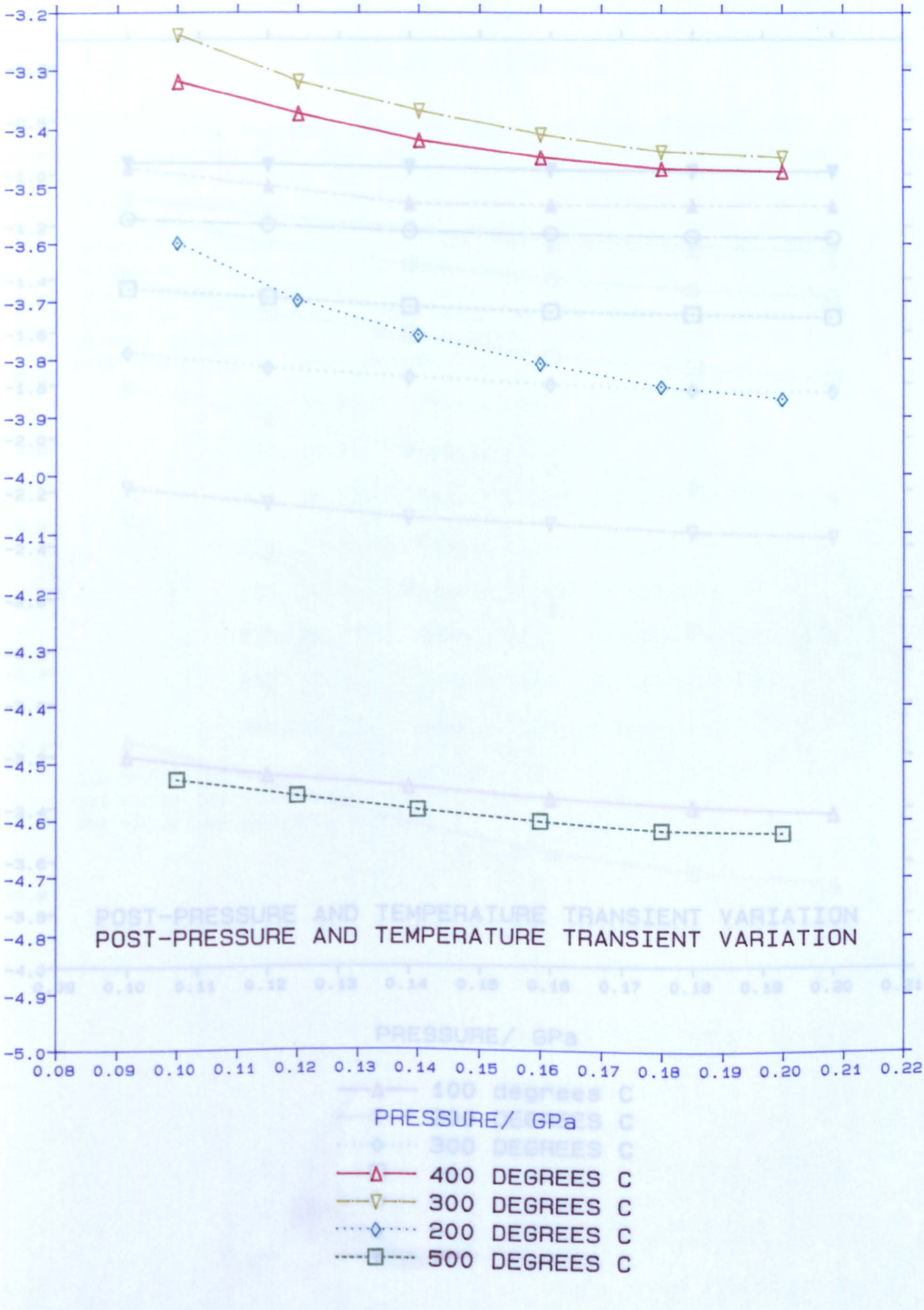


FIG. F8.68. GGRAN1PT PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATIONS AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE

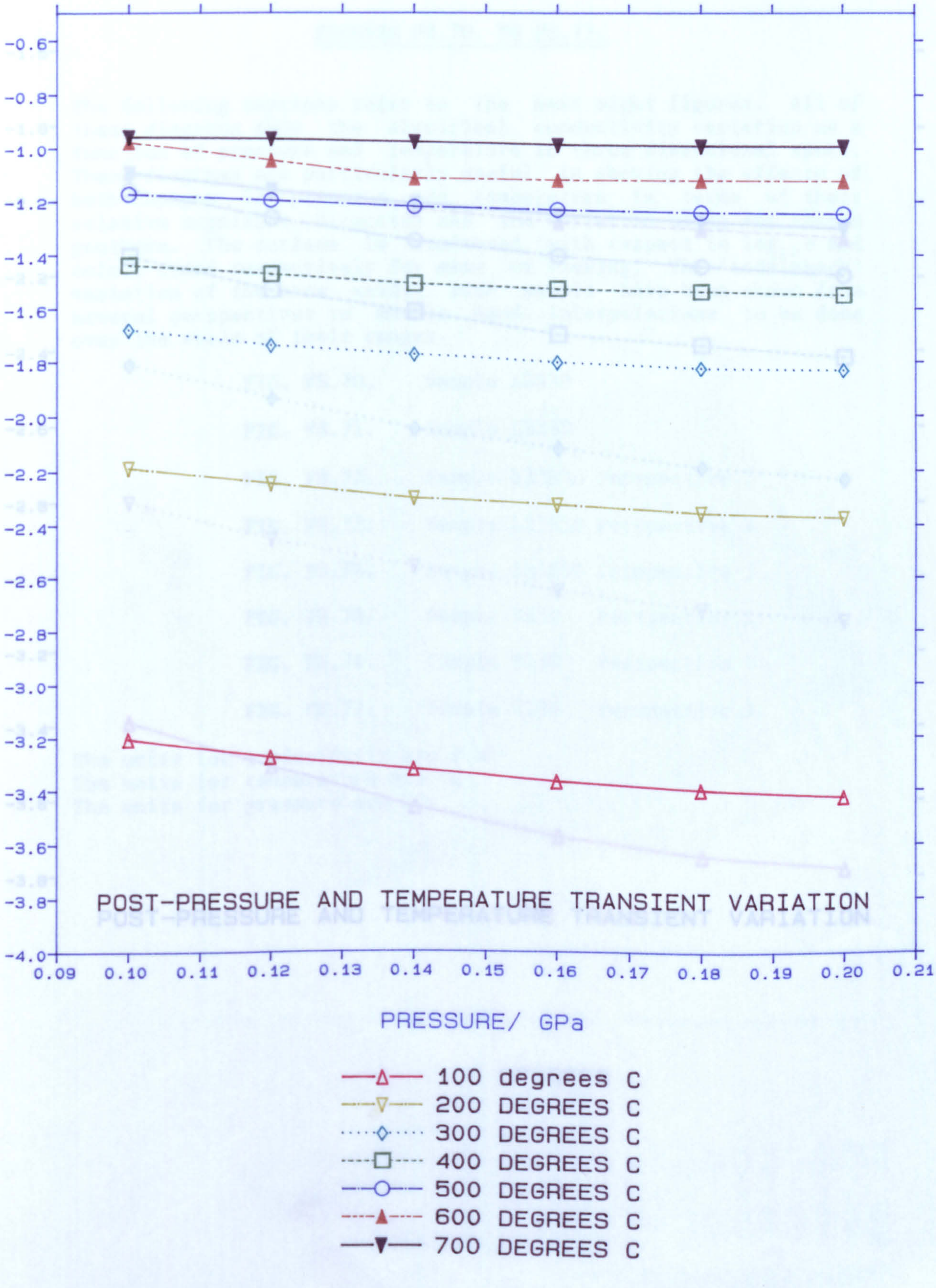
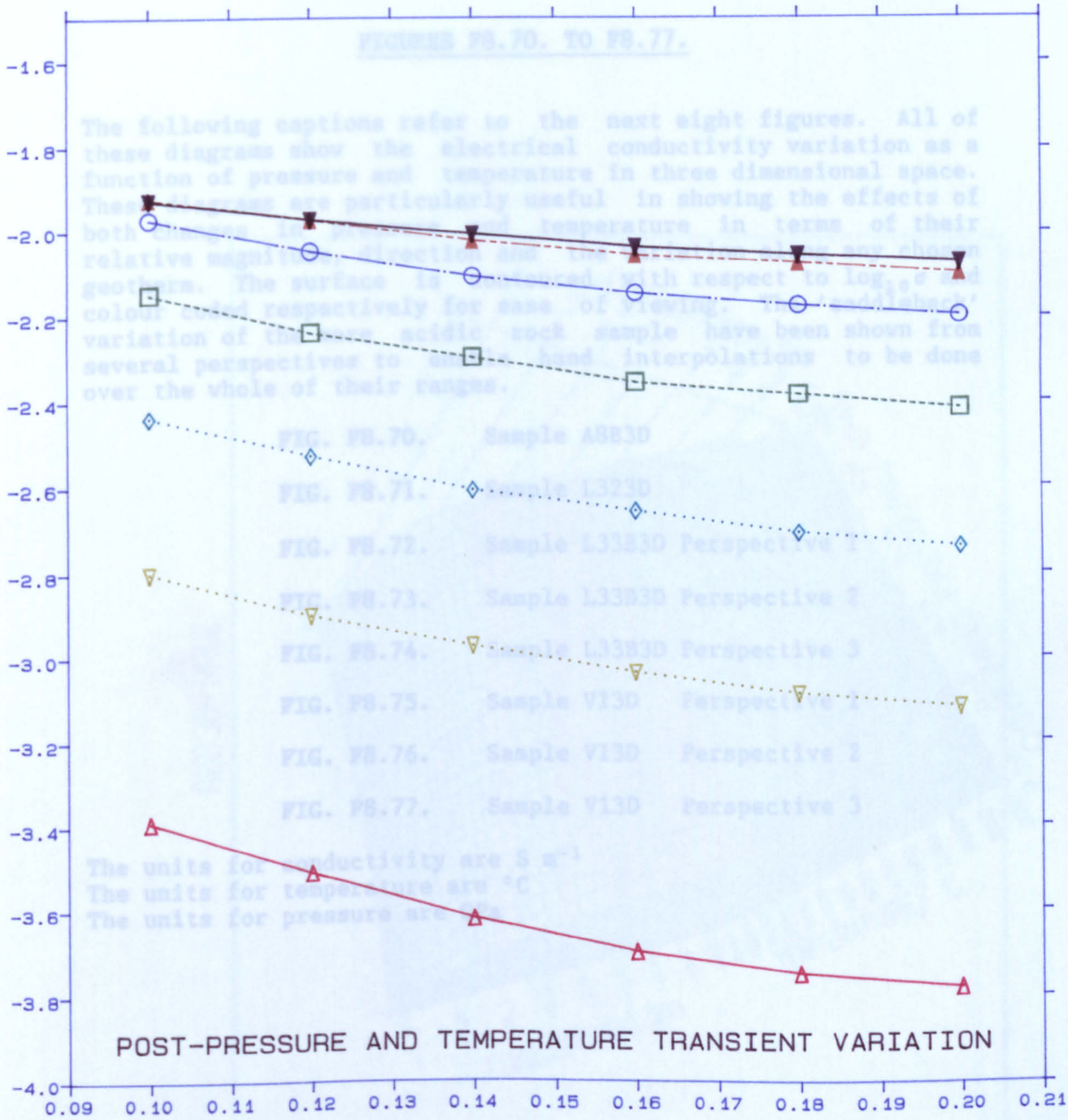


FIG. F8.69. GGRAN2PT PRESSURE VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATIONS AT VARIOUS TEMPERATURES AND 0.08 GPa PORE FLUID PRESSURE



FIGURES F8.70. TO F8.77.

The following captions refer to the next eight figures. All of these diagrams show the electrical conductivity variation as a function of pressure and temperature in three dimensional space. These diagrams are particularly useful in showing the effects of both changes in pressure and temperature in terms of their relative magnitude, direction and the variation along any chosen geotherm. The surface is contoured with respect to $\log_{10}\sigma$ and colour coded respectively for ease of viewing. The 'saddleback' variation of the more acidic rock sample have been shown from several perspectives to enable hand interpolations to be done over the whole of their ranges.

FIG. F8.70. Sample A8B3D

FIG. F8.71. Sample L323D

FIG. F8.72. Sample L33B3D Perspective 1

FIG. F8.73. Sample L33B3D Perspective 2

FIG. F8.74. Sample L33B3D Perspective 3

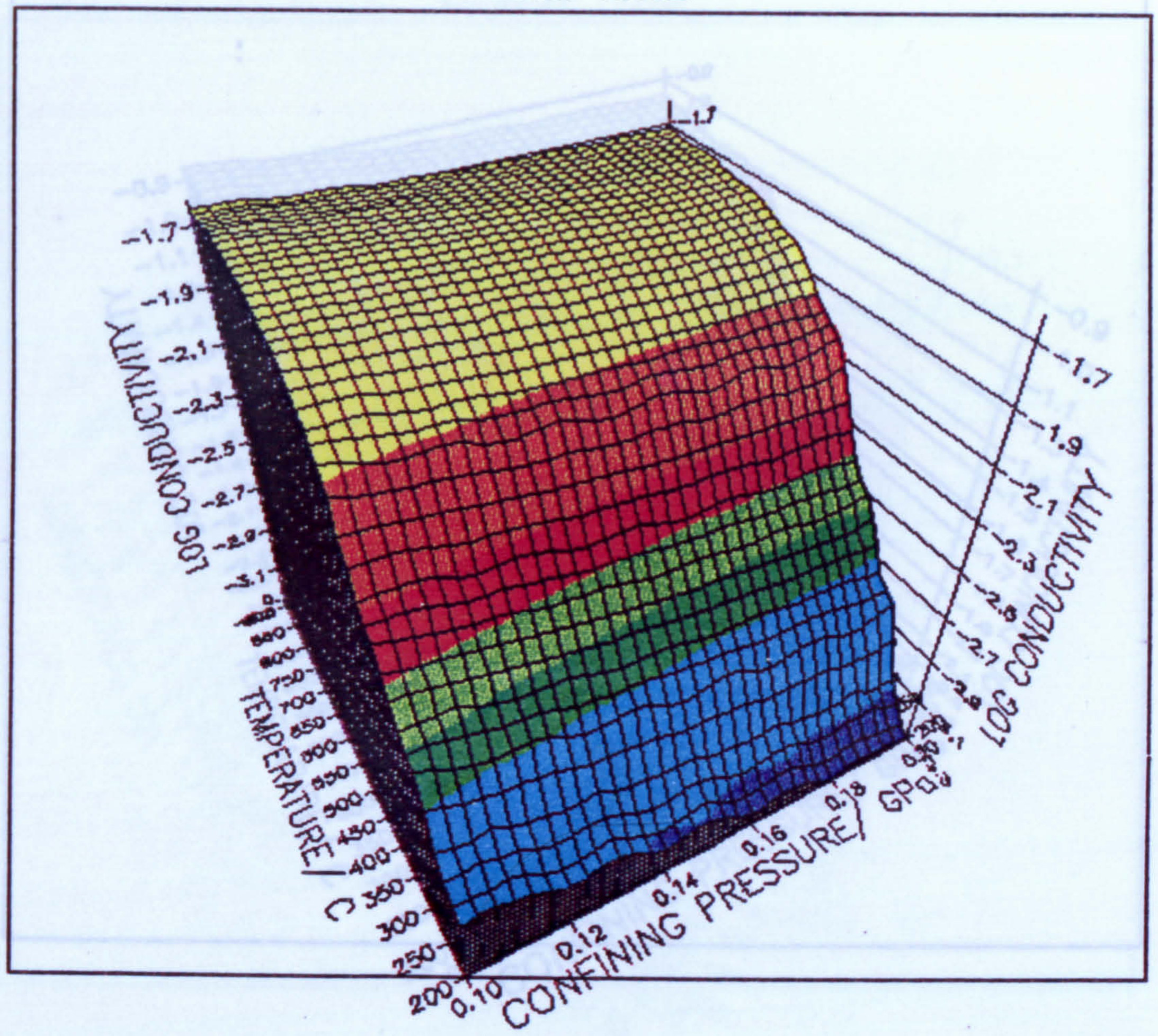
FIG. F8.75. Sample V13D Perspective 1

FIG. F8.76. Sample V13D Perspective 2

FIG. F8.77. Sample V13D Perspective 3

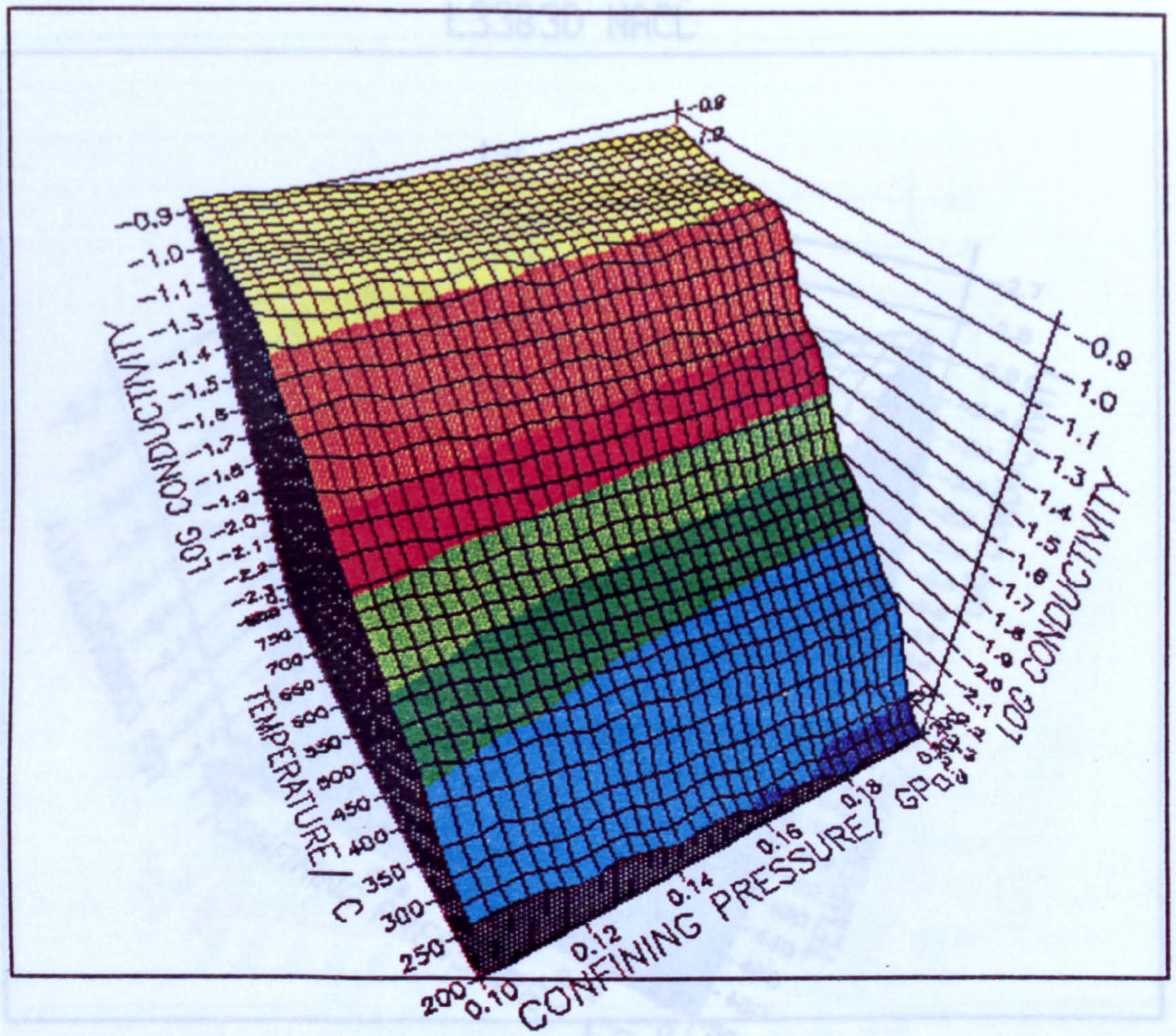
The units for conductivity are S m^{-1}
The units for temperature are °C
The units for pressure are GPa

A8B3D NACL



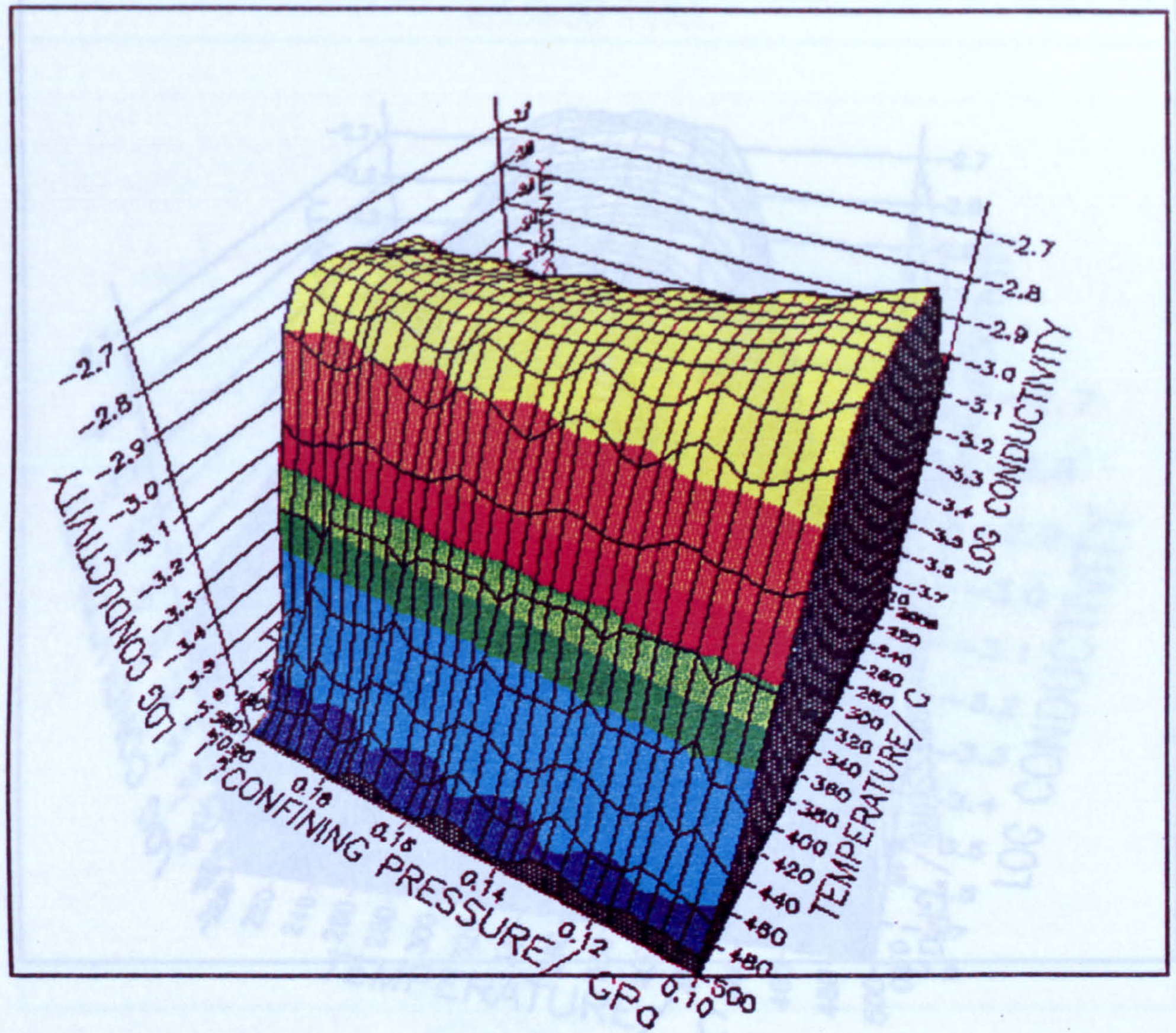
ABOVE	-1.6
-1.8	- -1.6
-1.9	- -1.8
-2.1	- -1.9
-2.2	- -2.1
-2.4	- -2.2
-2.5	- -2.4
-2.7	- -2.5
-2.8	- -2.7
-3.0	- -2.8
-3.2	- -3.0
BELOW	-3.2

L323D NACL



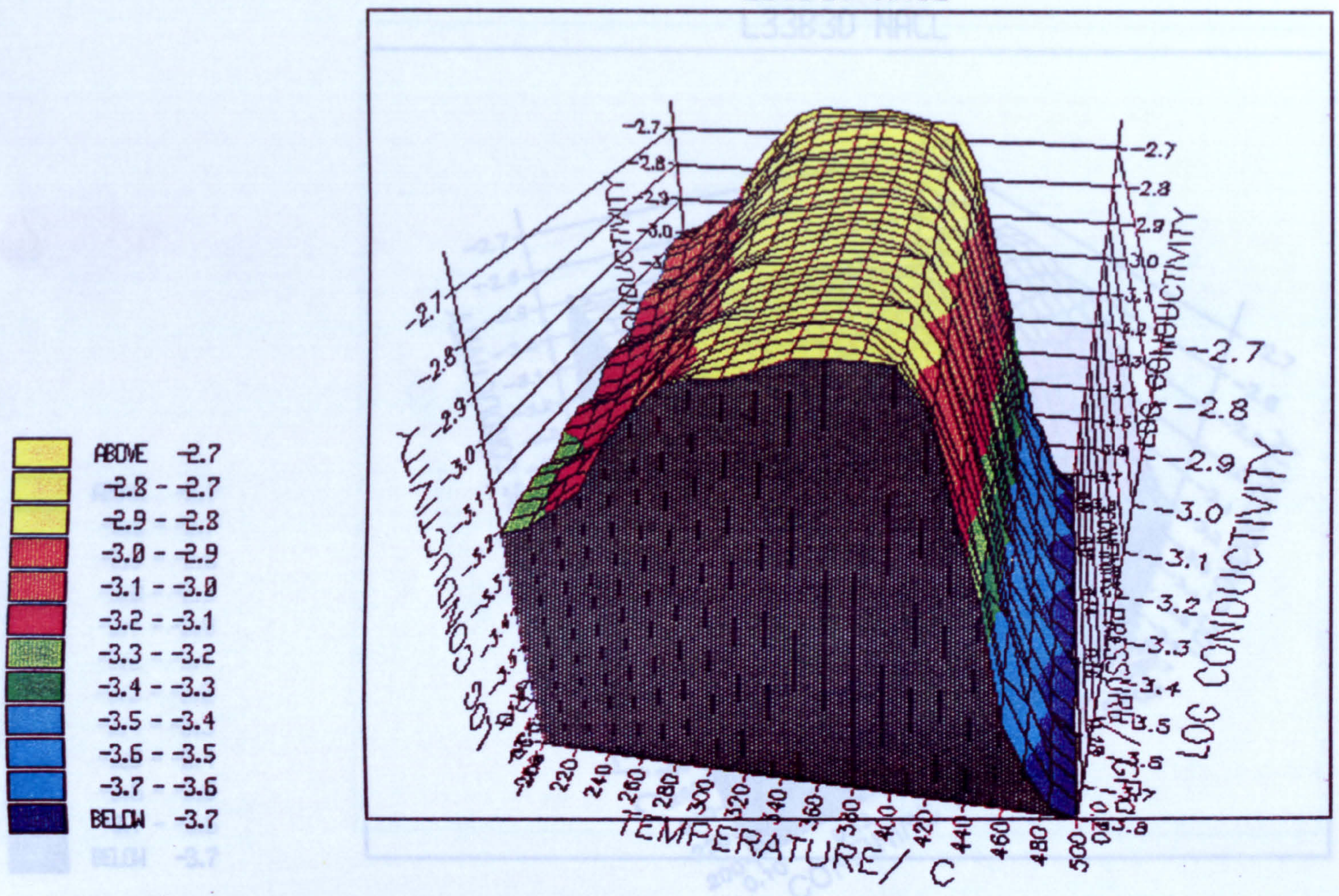
ABOVE	-0.8
-0.9	- -0.8
-1.1	- -0.9
-1.2	- -1.1
-1.4	- -1.2
-1.5	- -1.4
-1.6	- -1.5
-1.8	- -1.6
-1.9	- -1.8
-2.1	- -1.9
-2.3	- -2.1
BELOW	-2.3

L33B3D NACL

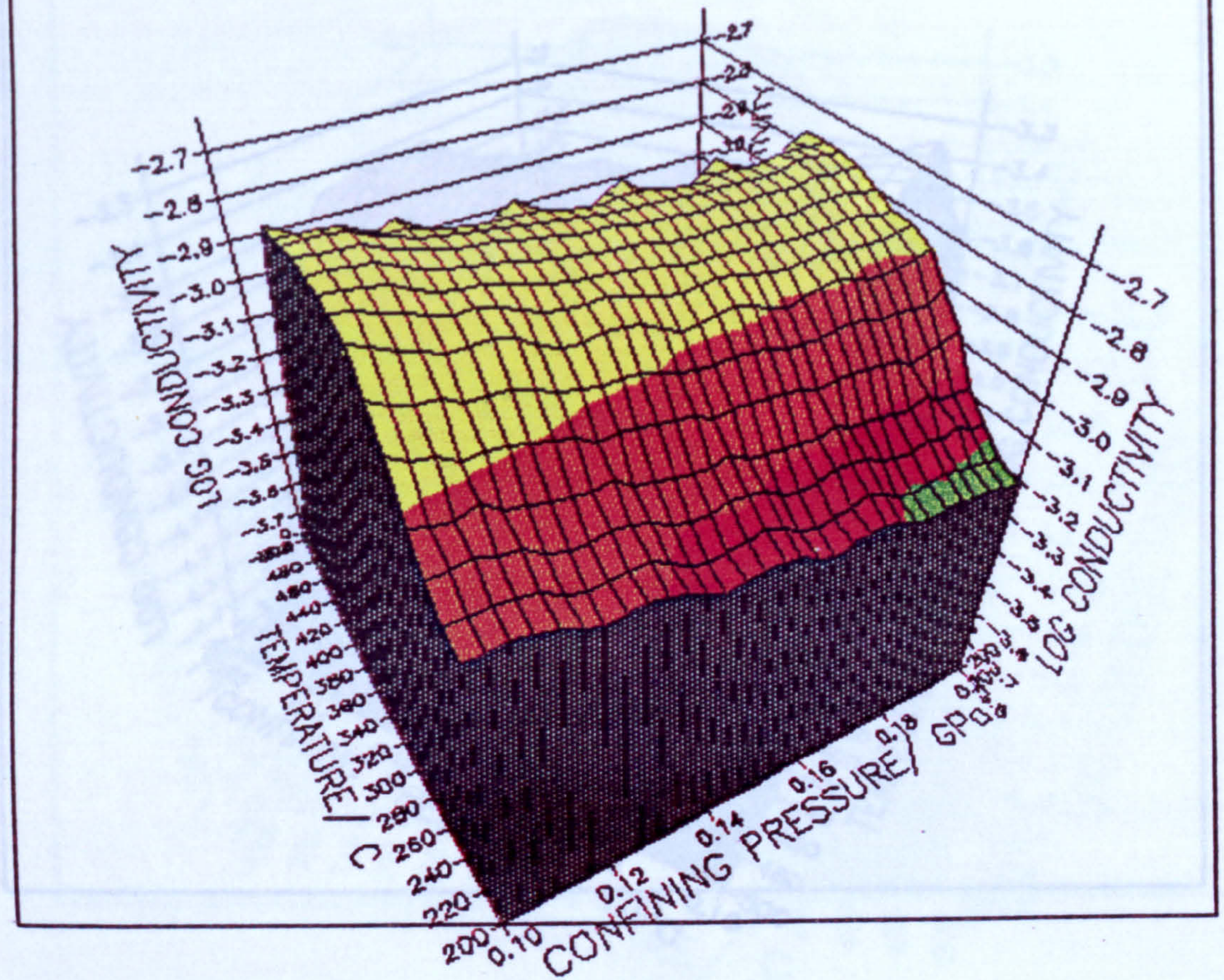


ABOVE	-2.7
-2.8	-2.7
-2.9	-2.8
-3.0	-2.9
-3.1	-3.0
-3.2	-3.1
-3.3	-3.2
-3.4	-3.3
-3.5	-3.4
-3.6	-3.5
-3.7	-3.6
BELOW	-3.7

L33B3D NACL

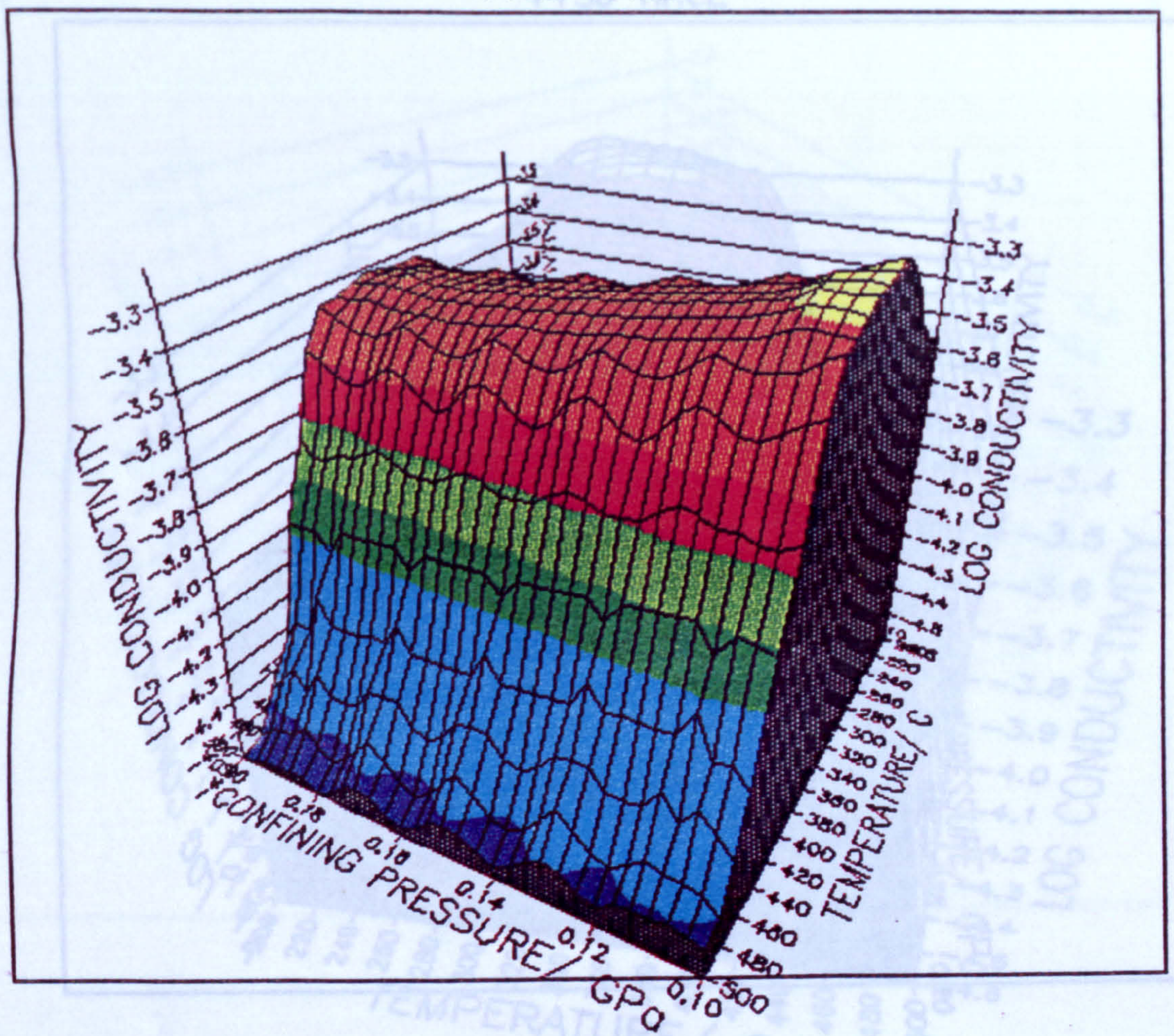

ABOVE -2.7
-2.8 - -2.7
-2.9 - -2.8
-3.0 - -2.9
-3.1 - -3.0
-3.2 - -3.1
-3.3 - -3.2
-3.4 - -3.3
-3.5 - -3.4
-3.6 - -3.5
-3.7 - -3.6
BELOW -3.7
LOG CONDUCTIVITY
-2.7
-2.8
-2.9
-3.0
-3.1
-3.2
-3.3
-3.4
-3.5
-3.6
-3.7
TEMPERATURE/ C
220
240
260
280
300
320
340
360
380
400
420
440
460
480
500
LOG CONDUCTIVITY
-2.7
-2.8
-2.9
-3.0
-3.1
-3.2
-3.3
-3.4
-3.5
-3.6
-3.7

L33B3D NACL



ABOVE -2.7
-2.8 - -2.7
-2.9 - -2.8
-3.0 - -2.9
-3.1 - -3.0
-3.2 - -3.1
-3.3 - -3.2
-3.4 - -3.3
-3.5 - -3.4
-3.6 - -3.5
-3.7 - -3.6
BELOW -3.7

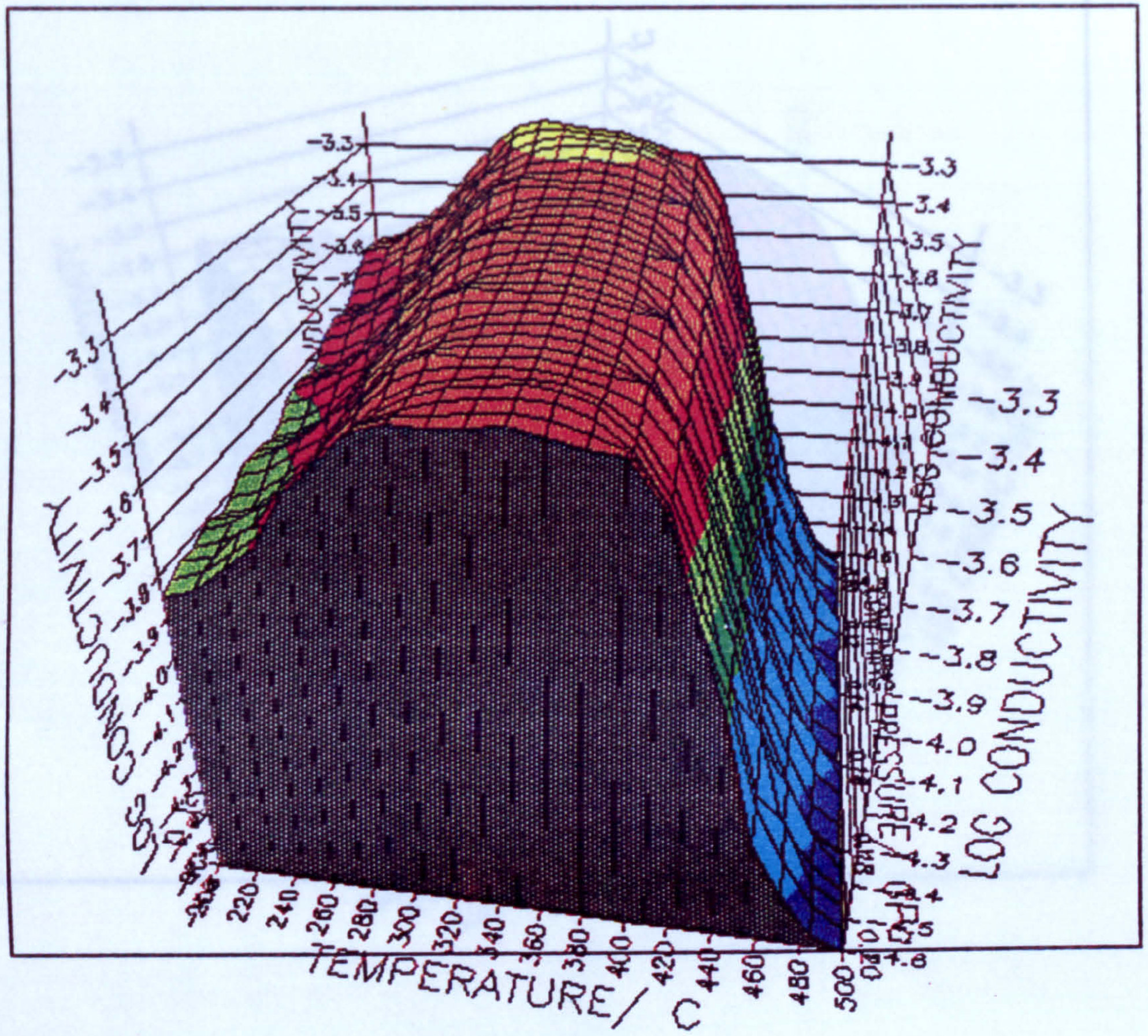
V13D NACL



ABOVE -3.0
-3.1 - -3.0
-3.3 - -3.1
-3.4 - -3.3
-3.6 - -3.4
-3.7 - -3.6
-3.9 - -3.7
-4.0 - -3.9
-4.2 - -4.0
-4.3 - -4.2
-4.5 - -4.3
BELOW -4.5

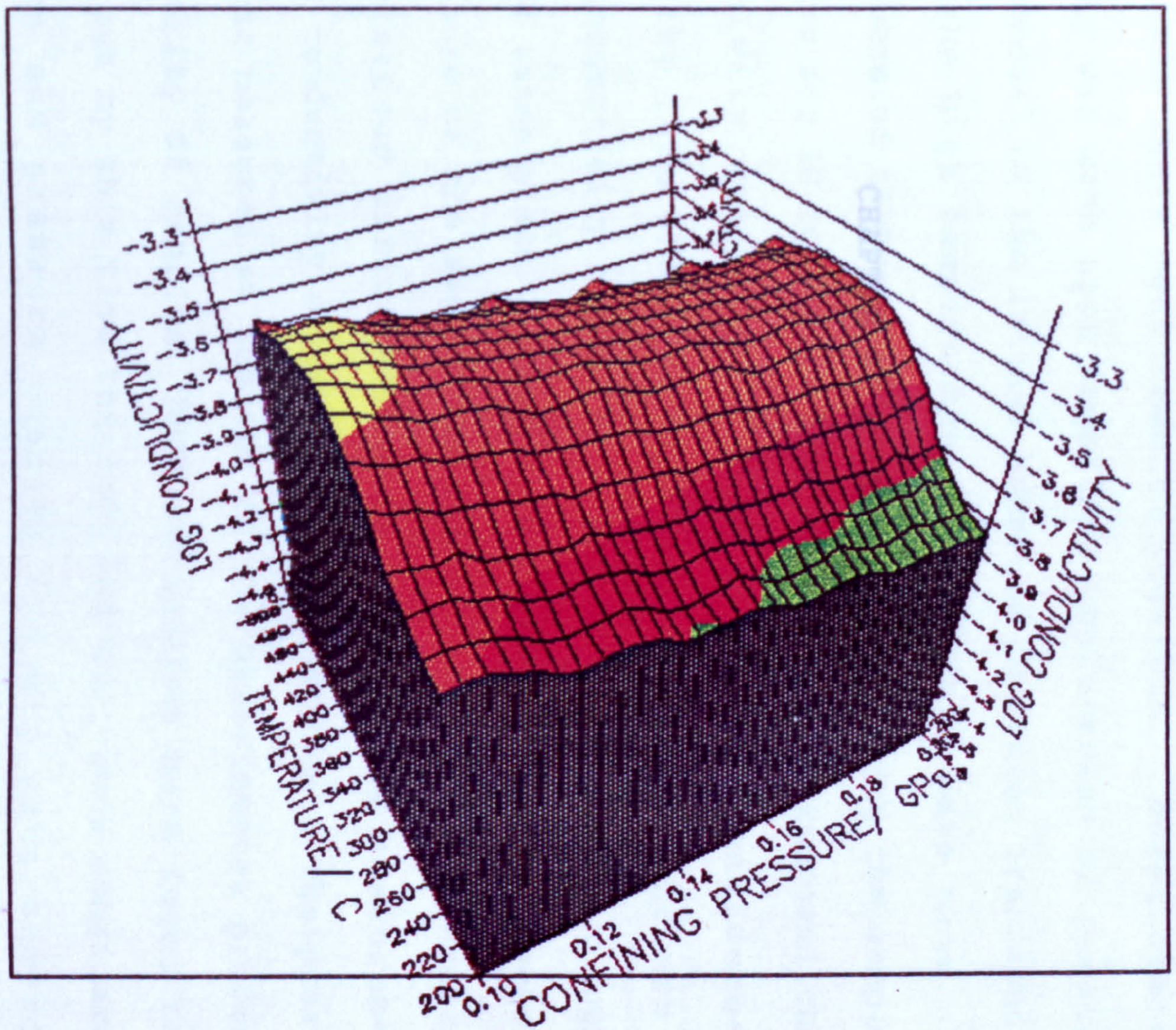
V13D NACL

V13D NACL



ABOVE	-3.0
-3.1	-3.0
-3.3	-3.1
-3.4	-3.3
-3.6	-3.4
-3.7	-3.6
-3.9	-3.7
-4.0	-3.9
-4.2	-4.0
-4.3	-4.2
-4.5	-4.3
BELOW	-4.5

V13D NACL



ABOVE	-3.0
-3.1	-3.0
-3.3	-3.1
-3.4	-3.3
-3.6	-3.4
-3.7	-3.6
-3.9	-3.7
-4.0	-3.9
-4.2	-4.0
-4.3	-4.2
-4.5	-4.3
BELOW	-4.5

CHAPTER 9 : DISCUSSION

9.1. THE EFFECT OF CONFINING PRESSURE UPON THE ELECTRICAL CONDUCTIVITY OF SATURATED ROCK SAMPLES

9.1.1 INTRODUCTION

Measurement of the confining pressure variation of conductivity at low temperatures was not a primary aim of this work as many studies of this have already been done (**Lee, 1977; Evans, 1980; Jackson, 1984;** etc). The order of the experimental procedure has ,however, made these measurements possible. Measurement of conductivity was made upon immediate application of pressure, after the removal of the initial pressure generated transient and after all the high temperature measurements were made. The pressure generated transient was removed by leaving the sample at high pressure for 10 to 12 hours by which time no temporal change of conductivity was noted. This state was considered to represent the case where all changes of microstructure and therefore conductivity caused by a particular change in applied pressure had taken place. There were further temporal changes if the temperature of the system was raised. Holding the sample at high temperatures during the main part of the measurements removed all conductivity effects from this source. The pressure variation was measured at the end of the experimental procedure in the majority of samples. Such variations were found to be repeatable but by this time the rock had undergone considerable temperature and pressure changes and may have undergone significant alteration from its pre-experimental state.

It has long been recognised that repeatable low temperature conductivity measurements can only be made after a sample has been cycled up to high confining pressures and left there for the

rock to reach a constant microstructure that is stable in such a confining pressure and temperature regime (**Lee, 1977; Jackson, 1984**). This has been discussed fully in Chapter 2 and can be seen in the pressure variations of figures F8.1. to F8.14. It has also been noted that a further transient conductivity drop occurs on subsequently leaving the sample at high pressure at a higher temperature (figures F8.27. to F8.38. and figures F8.1-F8.14), the magnitude of the effect being correlatable to the magnitude of the temperature increase (**Evans, 1980; Jackson, 1984**). The variation of conductivity during the process of the temperature activated transient drop has not often been measured.

The pore pressure was raised partly before holding the sample at raised temperature and confining pressures and subsequently increased to higher values for further high temperature and confining pressure measurements in most samples. Previously measurements have been thought to be reliable and truly representative of the rock in its **in situ** state after the removal of the pressure transient by cycling. Though typically significant, the subsequent drops taking place after temperature is raised have been ignored.

The monitoring and analysis of the characteristic relaxation time of the electrical conductivity while the sample was being held at high pressures and temperature during the transient drops may give further insight into the way that the rock microstructure responds to high temperature and pressure regimes. Such measurements were not made in this work as the nature and the

length of time of each experimental run required the operator to get some rest during the 100 or so hours that each run took to complete.

Most of the samples in this study were measured before prolonged exposure to high pressure, after this and after prolonged exposure to high pressures and temperatures. The initial confining pressure variation of electrical conductivity was determined by single conductivity measurements (figures F8.1. to F8.14.). The variation may not wholly be due to closure of cracks within the sample. **Lee (1977)** found that there was an anomalously high conductivity at very low confining pressures which he attributed to pore fluid conduction along the surface between the sleeve and the sample. He confirmed this by making measurements upon a PTFE sample. Such an experiment was also done with the new cell. It was found that the use of a guard-ring significantly reduced the the current leakage at very low confining pressures. The use of the guard-ring system reduced the apparent leakage conductivity by around 1.5 orders of magnitude. Above about 0.01 GPa the sleeve was compressed against the rock so tightly that the apparent conductivity due to leakage was two magnitudes less than conductivities typical of saturated rocks. The leakage current was beyond the measurement capabilities of the A.C. bridge instrument when the guard-ring was turned on and pressures were greater than 0.01 GPa. The very low surface leakage was attributed to the way in which the rock was sleeved. Removal of the sleeve after the experiment required the use of a very sharp knife. On removal, the inner surface of the sleeve showed a very fine reproduction of the granularity of the sample surface. The sleeve had been held in very close

correspondence to the sample surface thus reducing the surface conduction greatly.

9.1.2. TRANSIENT DROPS

A transient drop in conductivity at raised confining pressure and low temperature, as observed by Lee (**Lee, 1977**), was noted in all the samples (figures F8.1. to F8.14.). Subsequent post pressure transient conductivity variation with confining pressure (figures F8.1 to F8.14.) were determined by making repeated cycled measurements on increase and decrease of pressure. The post pressure variation showed little hysteresis and was repeatable providing the temperature was kept low and constant.

A further transient drop was found to occur after the sample was kept at raised pressures and raised temperatures. The post pressure and temperature transient response of conductivity to pressure is shown in figures F8.5. to F8.11. The measurements were made by making repeated cycled measurements of conductivity on raising and lowering of confining pressure at the end of the experimental procedure, after the sample had been subject to a long period at both high pressures and temperatures. The temperature transient was larger than the pressure transient in all the rocks studied.

The pressure transient is generally recognised as occurring due to slippage along grain boundaries and permanent closure of cracks and pores due to applied pressure. Most rocks respond relatively elastically for low confining pressure changes. The recovery period on removal of the pressure may, however, be very

large (Lee, 1977; Evans et al., 1982). As pressure is increased the rocks begin to behave inelastically and the compaction process becomes more permanent. The pore structures affected by the increase of pressure changes as the confining pressure is increased as described by **Walsh (1965)**. At low confining pressures it is the low aspect ratio cracks that close, and these tend to reopen again on removal of the confining pressure or application of pore pressure. At higher pressures all the LAR cracks have closed and the confining pressure begins to effect closure of the higher aspect ratio pores. The mechanism of this is unclear but if it includes grain slippage it may result in the permanent changes observed in rocks in their post-high pressure transient state. Subsequent release of pressure may result in the low aspect ratio cracks reopening completely but the high aspect ratio cracks attaining no or limited recovery. The result would be to alter the pore aspect ratio spectrum of the rock as it was cycled up to progressively higher confining pressures. Such an effect should be observable by measuring the mercury porosimetry of the sample between subsequent confining pressure changes and by acoustic monitoring and inversion of the V_p and V_s waves to give a pore aspect ratio spectrum or by measuring the bulk volume change in the sample during pressurisation.

The transient change takes place over a few hours for mid-crustal applied pressures and stops when the strength of the rock matrix is balanced out by the effective pressure acting on the rock. The high temperature and pressure transient is larger for larger temperature increases. It is an exacerbation of the previous pressure effect by high temperatures. Increasing the temperature of a rock increases the pore compressibility **(Kulikov, 1971)** such

that a rock that has previously reached an equilibrium state, in which the applied effective pressure is balanced by the rock strength, is disturbed. Pore compression is made easier by the raised temperatures and pores close until a new equilibrium is reached. The new equilibrium has a less porous, less connected pore network with a correspondingly lower conductivity. The decrease in conductivity due to this effect also has a timescale of hours.

The presence of a lubricating fluid may be an important factor influencing the speed with which rocks can readjust their microstructure to one which is stable in a higher pressure and temperature regime. I have not become aware of any work that has found that the presence of a fluid phase has enabled the electrical conductivity of a rock sample to reach a new equilibrium faster due to lubricated grain-slippage, although this may be the case. Alternatively, the presence of fluids saturating relatively unconnected pores may produce a high pore pressure upon the application of a confining pressure. Such a pore pressure would delay the compaction of the rock by keeping the pores open and therefore lengthen the amount of time required for the rock to come to a new stable state, lengthening the time taken for the high pressure and high temperature transients to occur. It must be noted that this argu ment only applies to relatively unconnected pores as any pore pressure rise in low aspect ratio connected pores due to confining pressure will diminish due to pore fluid flow out of the rock sample.

The rocks studied showed that the magnitude of the post-pressure

transient conductivity was about half an order of magnitude less than the initial conductivity and that the post-pressure and temperature conductivity was between a quarter and three quarters of an order of magnitude less than the post-pressure transient conductivity. The choice of which conductivity is judged to be the most representative of the situation in the crust is thus very important. Rocks in the middle and lower crust have had a very long term exposure to the high temperatures and pressures operating in their environment. It is therefore thought that any loss of conductivity due to compaction by combined high pressures and temperatures will have already occurred in them. The most representative measurements of the conductivity variation of crustal rocks with confining pressure are those taken after both the post-pressure and post-pressure and temperature transients have been removed by cycling and holding the sample at high temperatures and pressures.

The post-pressure and temperature transient measurements are, however, taken at low temperatures. An even more reliable measurement of the confining pressure variation of the electrical conductivity of rocks in the Earth would involve measurement at realistic deep crustal temperatures after the sample had been held for a while at these high temperatures and deep crustal pressures. This data has been obtained and is presented in section 8.4. The variations are shown parametrically as figures F8.63. to F8.69. and in 3 dimensional form in figures F8.70. to F8.77. These results will be further discussed in section 9.4.

It has been recognised that there may be differences between the values of measured conductivity before and after being held at

high temperatures may also be due to chemical and phase changes within the rock minerals.

9.1.3. THE SHAPE OF THE PRESSURE VARIATION

The shape of the apparent electrical conductivity versus confining pressure variation depends upon the crack and pore aspect ratio spectrum of the rock **(Lee, 1977)**. As pressure increases there is a progressive cutting off of pore conduction pathways due to the closure of cracks. The geometry of the crack and the strength of the rock will both control the pressure at which closure takes place. The smaller the crack aspect ratio or rock strength the lower will be the pressure at which the crack closes. All the samples studied showed a conductivity variation that can be expressed by the following equation;

$$\sigma^* = aP_c^{-b} \qquad \text{E9.1.}$$

or $$\log_{10} \sigma^* = \log_{10} a - b \log_{10} P_c \qquad \text{E9.2.}$$

The values of the constants a and b are shown in table T9.1.

TABLE T9.1. EMPIRICAL PRESSURE VARIATION PARAMETERS		
SAMPLE	$X10^{-3}$ S m^{-1} a	b
BASIC AMPHIBOLITE		
IVD1	2.51,1.99	0.85,0.90
IVD2 DD	0.32,0.22	10.5,1.15
BASIC GRANULITE		
IVE1	5.25,3.16	0.97,1.05
IVE2 DD	0.93,0.56	1.30,1.30
ORTHOPYROXENITE		
MER21	0.35,0.22,0.06	0.55,0.40,0.27
MER22 DD	0.058,0.033,0.018	0.50,0.55,0.55
AMPHIBOLITE GNEISS		
A8B	0.91,0.38,0.17	0.60,0.48,0.38
GRANODIORITE		
L33B	2.51,1.00,0.19	0.57,0.51,0.42
AMPHIBOLITE GNEISS		
L32	8.91,3.16,0.63	0.45,0.42,0.25
GRANITE		
V1	0.87,0.30,0.10	0.62,0.48,0.39
GRAPHITE GRANULITE		
GGRAN1	0.95,0.50,0.14	0.43,0.30,0.29
GGRAN2	0.63,0.28	0.73,0.65
GGRAN3 DRY	0.0050,0.0201	-0.80,-0.48
GGRAN4 DRY	0.0029,0.0035	-0.17,-0.13

Saturated with 0.5MNaCl solution unless otherwise stated: DD = double distilled water.

The first figure in each column refers to the initial variation, the second to the post-pressure transient variation and the third, if present, to the post-temperature transient variation.

The rate of variation of the post-transient values can be seen to be less than the initial variation in most cases by eye from the figures. It can also be seen from the data in the above table that the post-pressure transient and post-temperature transient values of b in general become progressively smaller. This can be attributed to the permanent loss of pore space upon the application of pressure and then further upon the application of temperature at high pressures and is consistent with the idea of pressure and temperature transient drops. It will also be noted

from the figures that the temperature transient drop tends to be larger than the pressure transient drop.

Table T9.1. does a similar job to a graph comparing the relative reductions of conductivity with pressure. Figures F9.1. to F9.3. show the comparison of the initial variations, post-pressure transient variations and post-temperature transient variations respectively. The acidic rocks are shown as broken lines for ease of comparison. It can be seen that there is no generic difference between the conductivity variations with pressure for acidic and metabasic rocks studied. This indicates that if there is an additional conduction mechanism acting in metabasic rocks it is relatively unaffected by the increase in pressure.

The more significant point, and one that is lost in the normalised figures F9.1. to F9.3. but can be noticed in table T9.1. is that the metabasic rocks tend to have initial conductivities larger than the acidic rocks. This indicates that an additional conduction mechanism is operating in such rocks. Direct comparison is hindered by the difference in porosities, but by comparing rocks with similar porosities judgements can be made. The basic granulite (IVE1) and the basic amphibolite (IVD1) from the Ivrea Zone and the amphibolite gneiss (L32) have the highest initial conductivities. The samples with the lowest conductivities were the granite (V1) and the orthopyroxenite (MER21). These samples also had lower porosities which may partially explain the lower values.

The difference of conductivities can be explained in terms of

FIGURE F9.1. THE VARIATION OF CONDUCTIVITY NORMALISED TO THAT AT ATMOSPHERIC PRESSURE WITH CONFINING PRESSURE FOR DIFFERENT ROCKS UNDERGOING AN INITIAL PRESSURE VARIATION

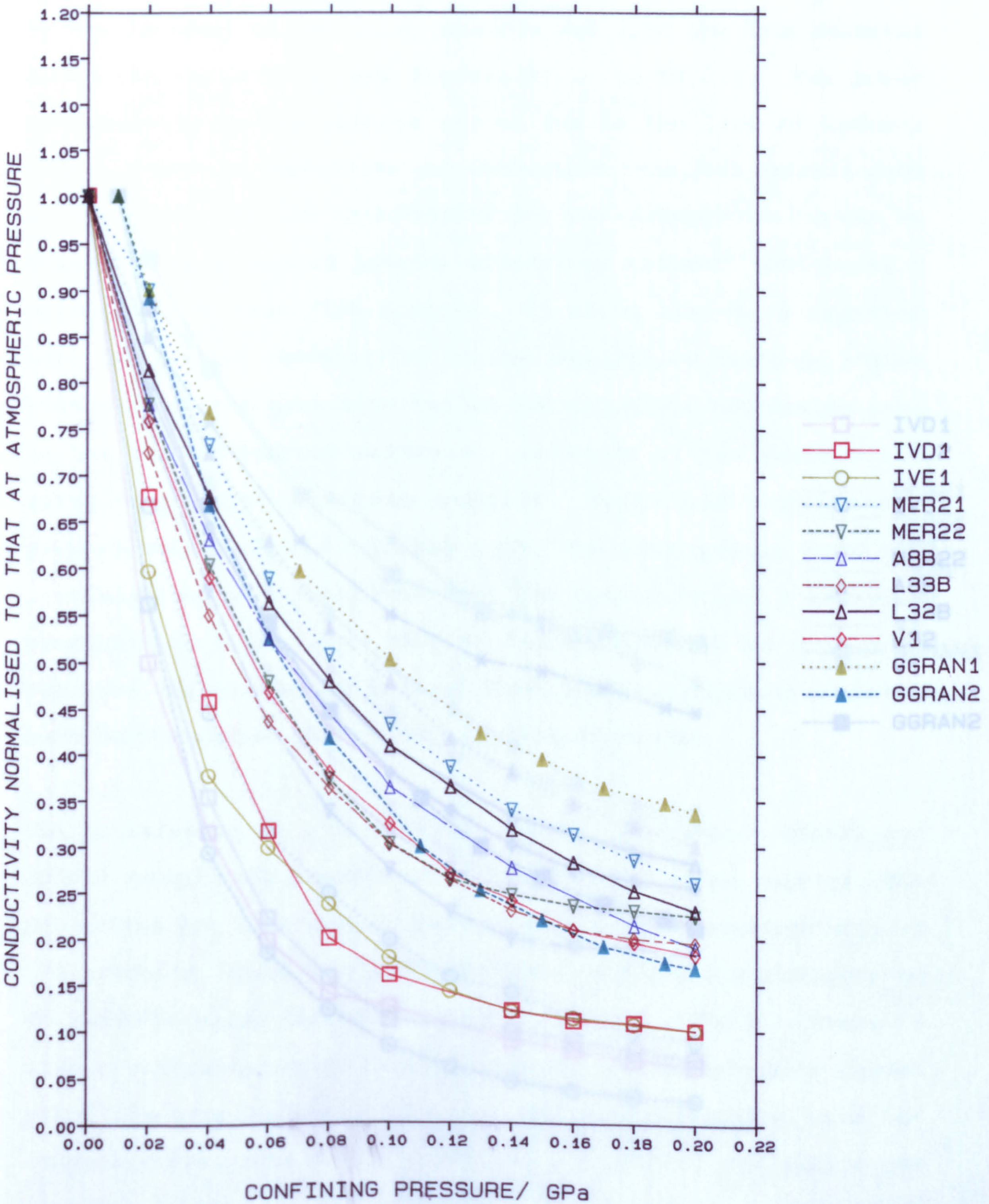


FIGURE F9.2. THE VARIATION OF CONDUCTIVITY NORMALISED TO THAT AT ATMOSPHERIC PRESSURE WITH CONFINING PRESSURE FOR DIFFERENT POST-PRESSURE TRANSIENT ROCKS

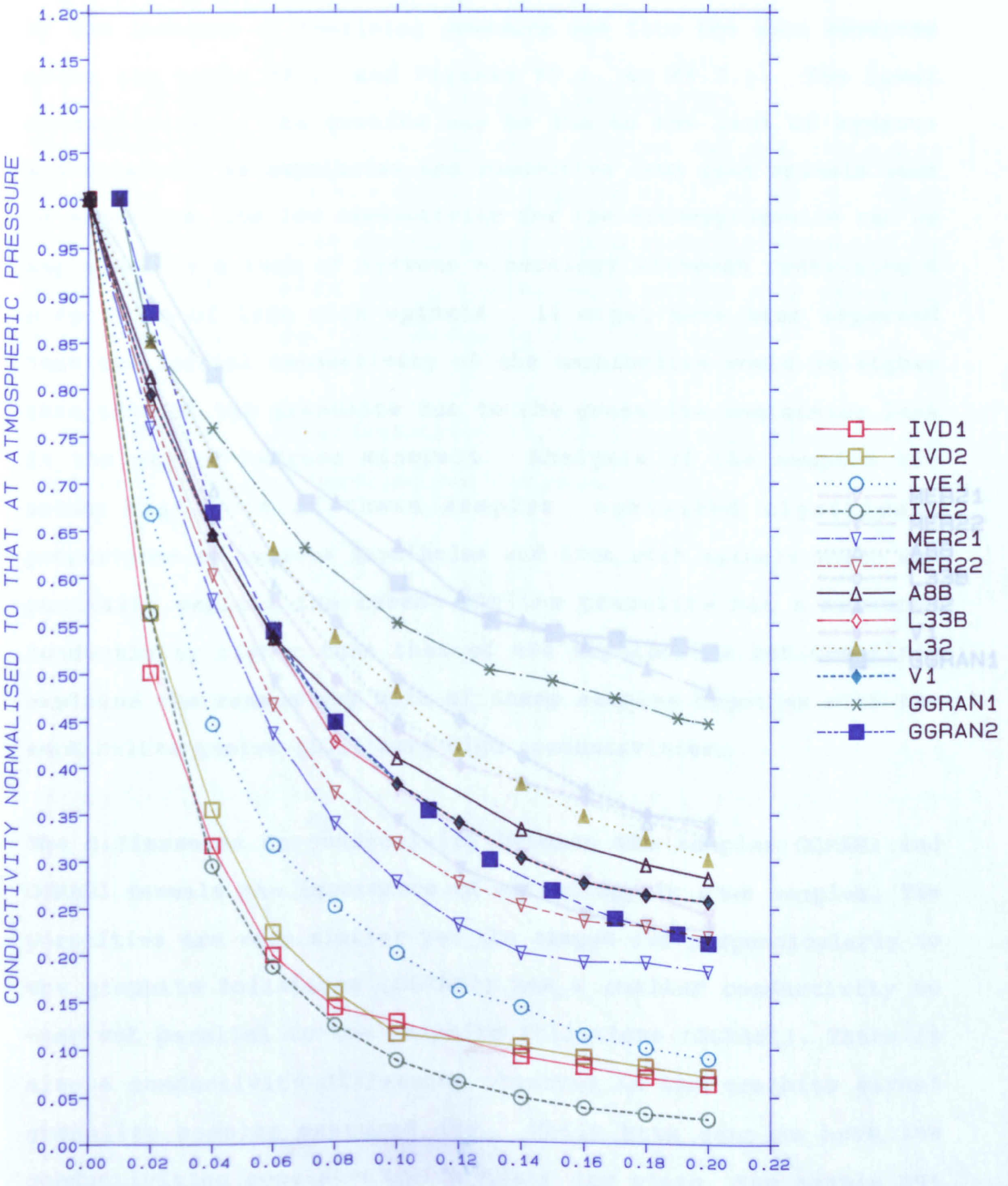
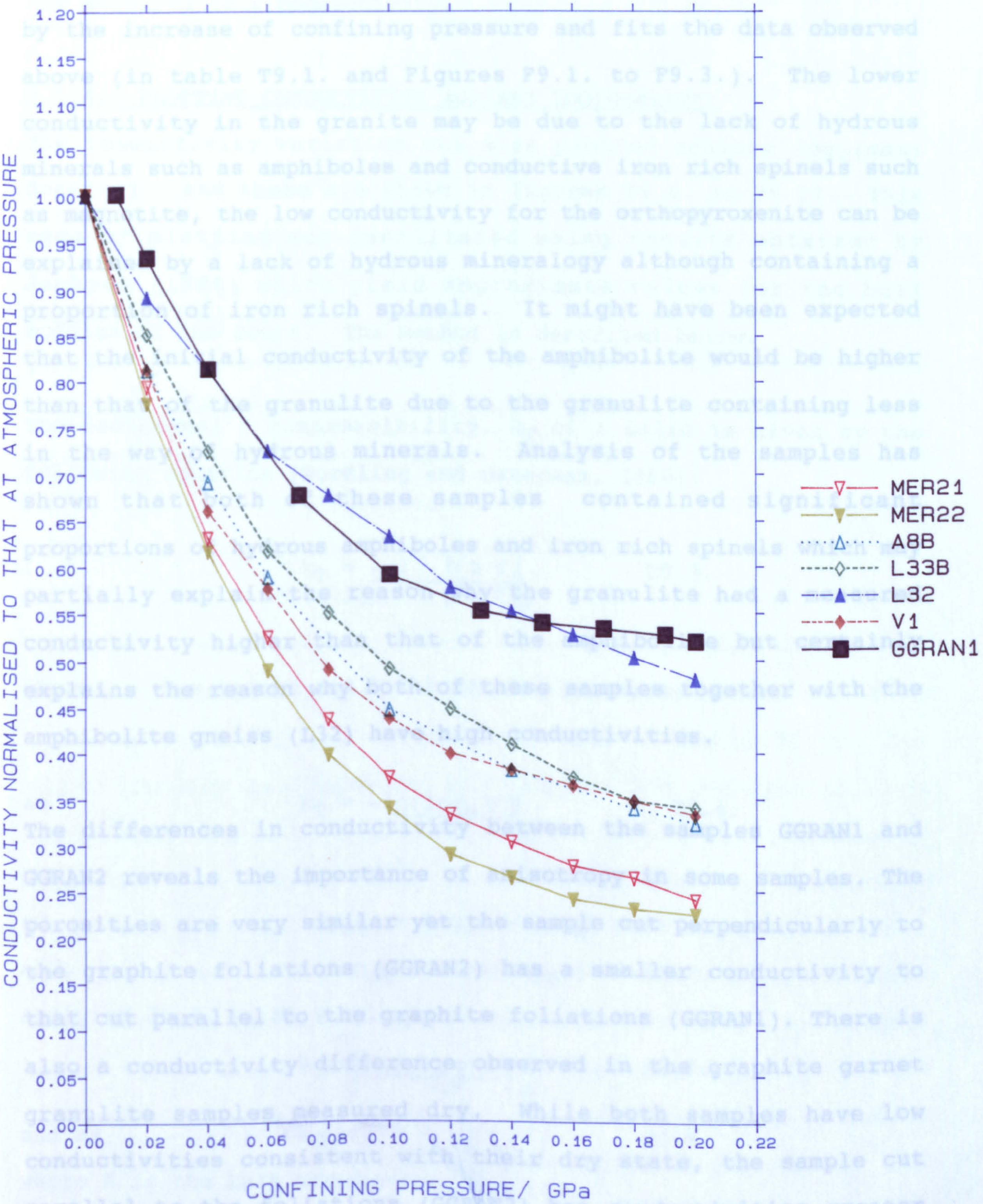


FIGURE F9.3. THE VARIATION OF CONDUCTIVITY NORMALISED TO THAT AT ATMOSPHERIC PRESSURE WITH CONFINING PRESSURE FOR DIFFERENT POST-TEMPERATURE TRANSIENT ROCKS



additional conduction occurring through solid state mineral conduction. Such conduction is easier in hydrous rocks, rocks with large cations and rocks containing iron-rich spinels such as magnetite. This type of conduction is not greatly affected by the increase of confining pressure and fits the data observed above (in table T9.1. and Figures F9.1. to F9.3.). The lower conductivity in the granite may be due to the lack of hydrous minerals such as amphiboles and conductive iron rich spinels such as magnetite, the low conductivity for the orthopyroxenite can be explained by a lack of hydrous mineralogy although containing a proportion of iron rich spinels. It might have been expected that the initial conductivity of the amphibolite would be higher than that of the granulite due to the granulite containing less in the way of hydrous minerals. Analysis of the samples has shown that both of these samples contained significant proportions of hydrous amphiboles and iron rich spinels which may partially explain the reason why the granulite had a measured conductivity higher than that of the amphibolite but certainly explains the reason why both of these samples together with the amphibolite gneiss (L32) have high conductivities.

The differences in conductivity between the samples GGRAN1 and GGRAN2 reveals the importance of anisotropy in some samples. The porosities are very similar yet the sample cut perpendicularly to the graphite foliations (GGRAN2) has a smaller conductivity to that cut parallel to the graphite foliations (GGRAN1). There is also a conductivity difference observed in the graphite garnet granulite samples measured dry. While both samples have low conductivities consistent with their dry state, the sample cut parallel to the foliations (GGRAN3) has conductivities greater

than the one cut perpendicular to the graphite foliations (GGRAN4). This demonstrates that anisotropy in rocks with a well defined texture should be considered as a mechanism for modifying the measured rock conductivity.

9.1.4. PLOTTING CONDUCTIVITY AGAINST LOG(DENSITY)

The conductivity variation was also plotted against $\log_e$(mass density) and these are shown in figures F9.4. to F9.17. This mode of plotting was facilitated using results obtained by Jackson (1984) which yield approximate values for the bulk modulus of the rocks. The method is described below.

The isothermal compressibility, k_T of a solid is given by the following equation (Nordling and Osterman, 1980):

$$k_T = \frac{-1}{V} \cdot \left|\frac{(\partial V)}{(\partial P)}\right|_T \qquad \text{E9.3.}$$

where; V = volume and P = confining pressure,

$$\text{so} \qquad k_T = -\left|\frac{d(\log_e V)}{dP}\right|_T \qquad \text{E9.4.}$$

$$\text{now} \qquad \rho = mV^{-1} \qquad \text{E9.5.}$$

$$\text{so} \qquad \log_e \rho = \log_e m - \log_e V \qquad \text{E9.6.}$$

$$\text{and} \qquad d(\log_e \rho) = -d(\log_e V) \qquad \text{E9.7.}$$

Applying E9.7. to E9.4. gives;

$$k_T = \left|\frac{d(\log_e \rho)}{dP}\right|_T \qquad \text{E9.8.}$$

$$\text{and as} \qquad k_T = K^{-1} \qquad \text{E9.9.}$$

where K is the bulk modulus

so $$K = \left| \frac{dP}{d(\log_e \rho)} \right|_T \qquad \text{E9.10.}$$

The P-wave and S-wave elastic velocities through the body of an elastic solid are given by:

$$V_P = \left\{ \frac{K + 4/3\mu}{\rho} \right\}^{1/2} \qquad \text{E9.11.}$$

$$V_S = \left\{ \frac{\mu}{\rho} \right\}^{1/2} \qquad \text{E9.12.}$$

where μ = shear modulus
ρ = body density

Jackson (1984) has made seismic velocity measurements on rocks similar to those used in this study. The average P and S-wave velocities were calculated from these results for three different types of rock;

- amphibolites
- granulite gneisses
- granites

These results are used in combination with equations E9.11. and E9.12. to generate average values of the bulk modulus of these types of rock. These values are given in table T9.2. The combination of the relevant bulk modulus for the type of rock under consideration, equation E9.10. and the initial density enabled the independent axis to be transformed from pressure to $\log_e$(density). The graphs are no more than linear isomorphic mappings with the bulk modulus acting as the scaling factor. The significance of this type of plotting is that the double logarithmic axes demonstrate that the conductivity varies linearly with density whereas the raw confining pressure curves have linear log(conductivity) with confining pressure and therefore exponentially decreasing conductivity with pressure.

The method also gives insight into how the closure of different types of pore affects the conductivity per unit pore volume loss.

TABLE T9.2. DENSITY TRANSFORMATION DATA				
ROCK SAMPLE	AVERAGE V_p km s^{-1}	AVERAGE V_s km s^{-1}	SHEAR MOD $\mu X10^3$ kg m^{-2} s^{-1}	BULK MOD $KX10^3$
AMPHIBOLITE	6.9	3.9	46.54	83.63
GRANULITE GNEISSES	6.65	3.6	39.92	82.98
GRANITE AND GRANODIORITE	6.2	3.45	30.95	58.67

Changes in the gradient of the $\log_{10}$(conductivity) against $\log_{10}$(density) graph show the extent to which a certain reduction of porosity, due to confining pressure compaction, affects the conductivity of the bulk of the rock. The effect on electrical conductivity due to a reduction of porosity is dependent upon the pore parameters. The rate of reduction of conductivity will be dependent upon the extent to which the particular type of porosity, being reduced at that pressure, has control over the whole rock conductivity. If the pressure is such that pores of a certain aspect ratio are being preferentially closed, and that particular topology of crack is important to the bulk rock conductivity, due to its geometry producing good pore interconnectivity, for instance, then the the log/log graph would be expected to have a greater negative gradient than if the pores were not of such a type. Factors that enable an individual pore to influence the sensitivity with which the rock conductivity changes with confining pressure per unit pore volume include the following;

FIG. F9.4. IVD1DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

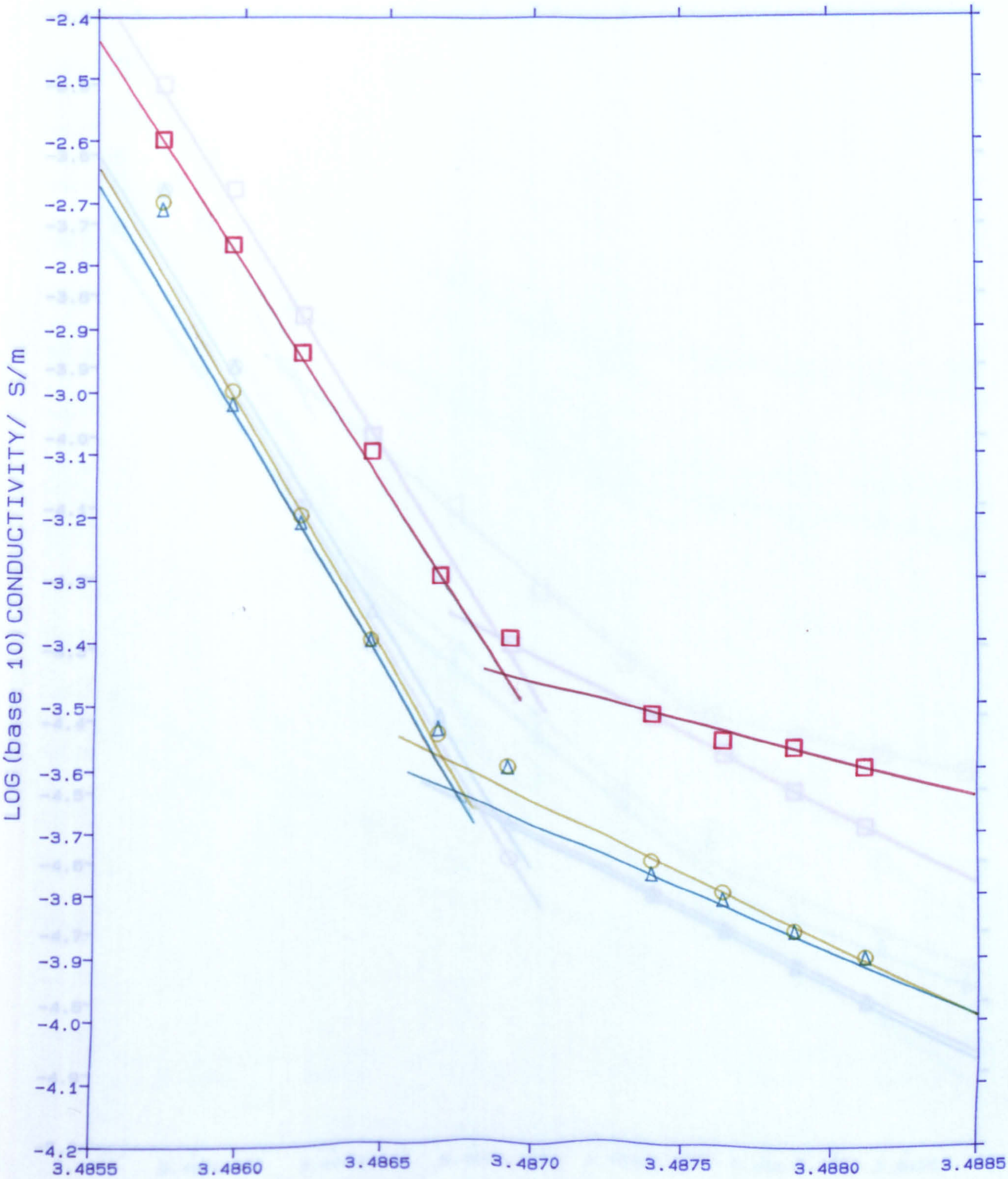


FIG. F9.5. IVD2DEN CONDUCTIVITY/DENSITY VARIATION FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

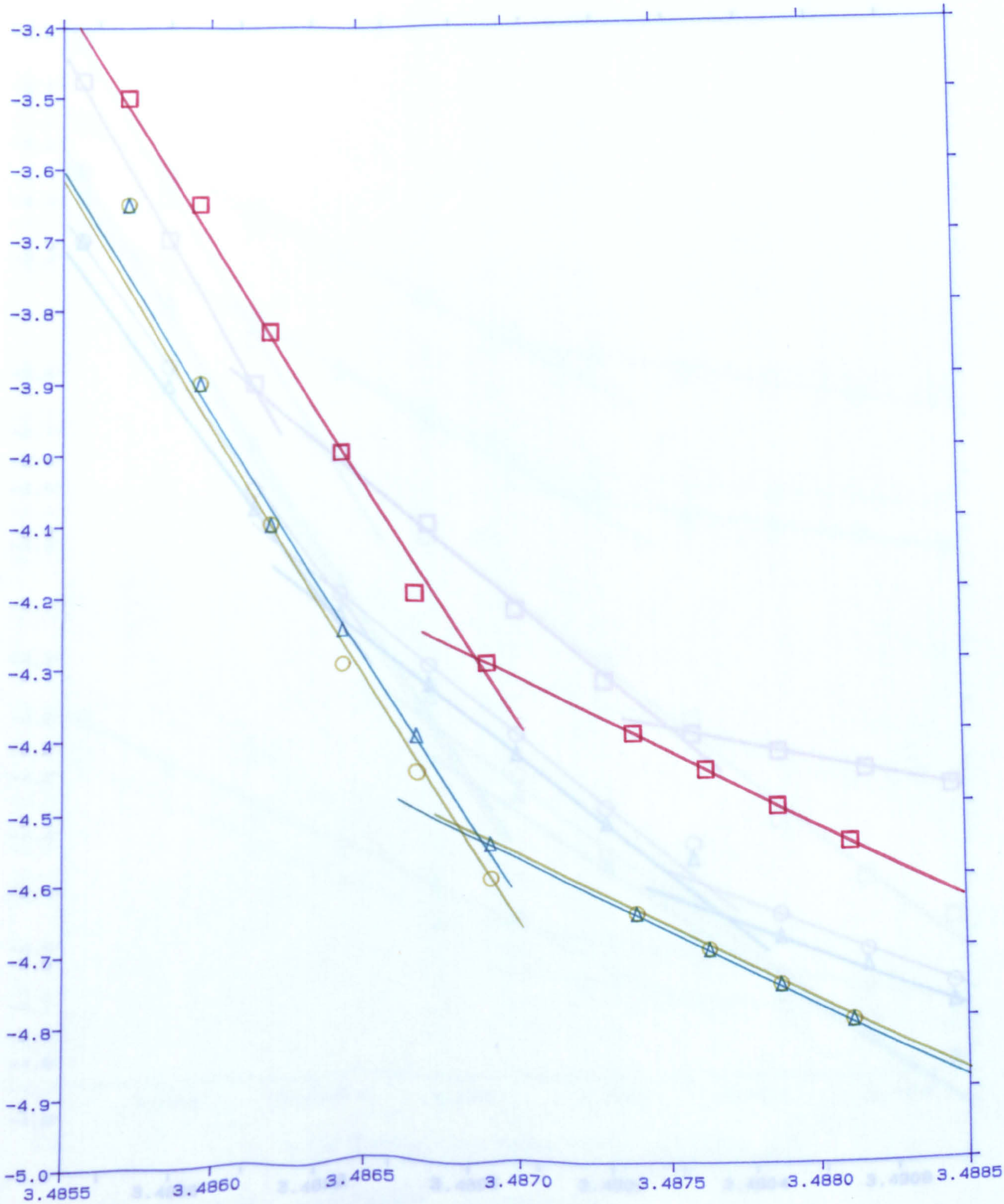


FIG. F9.6. IVE1DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

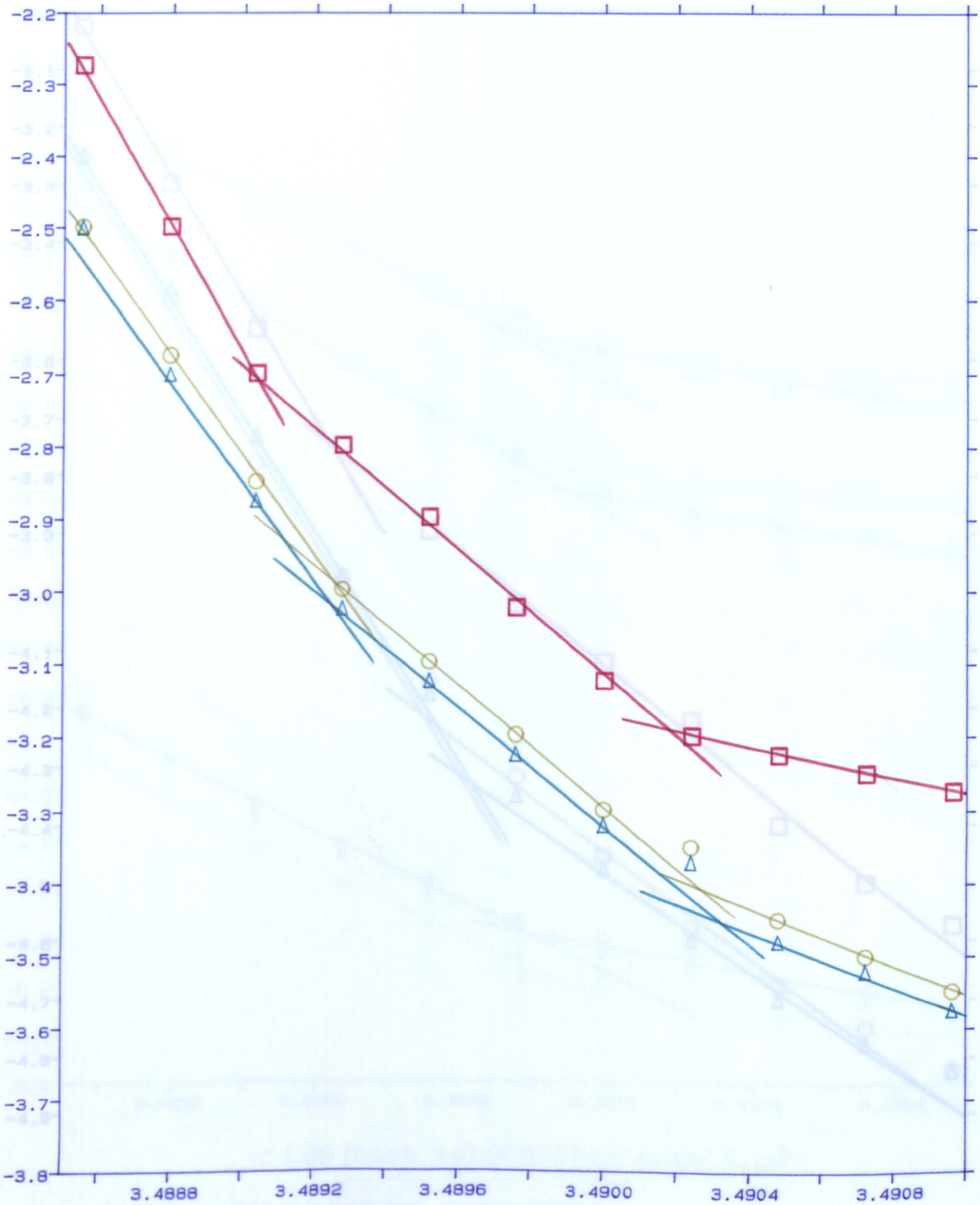


FIG. F9.7. IVE2DEN CONDUCTIVITY/DENSITY VARIATION FOR DISTILLED WATER SATURATED BASIC GRANULITE

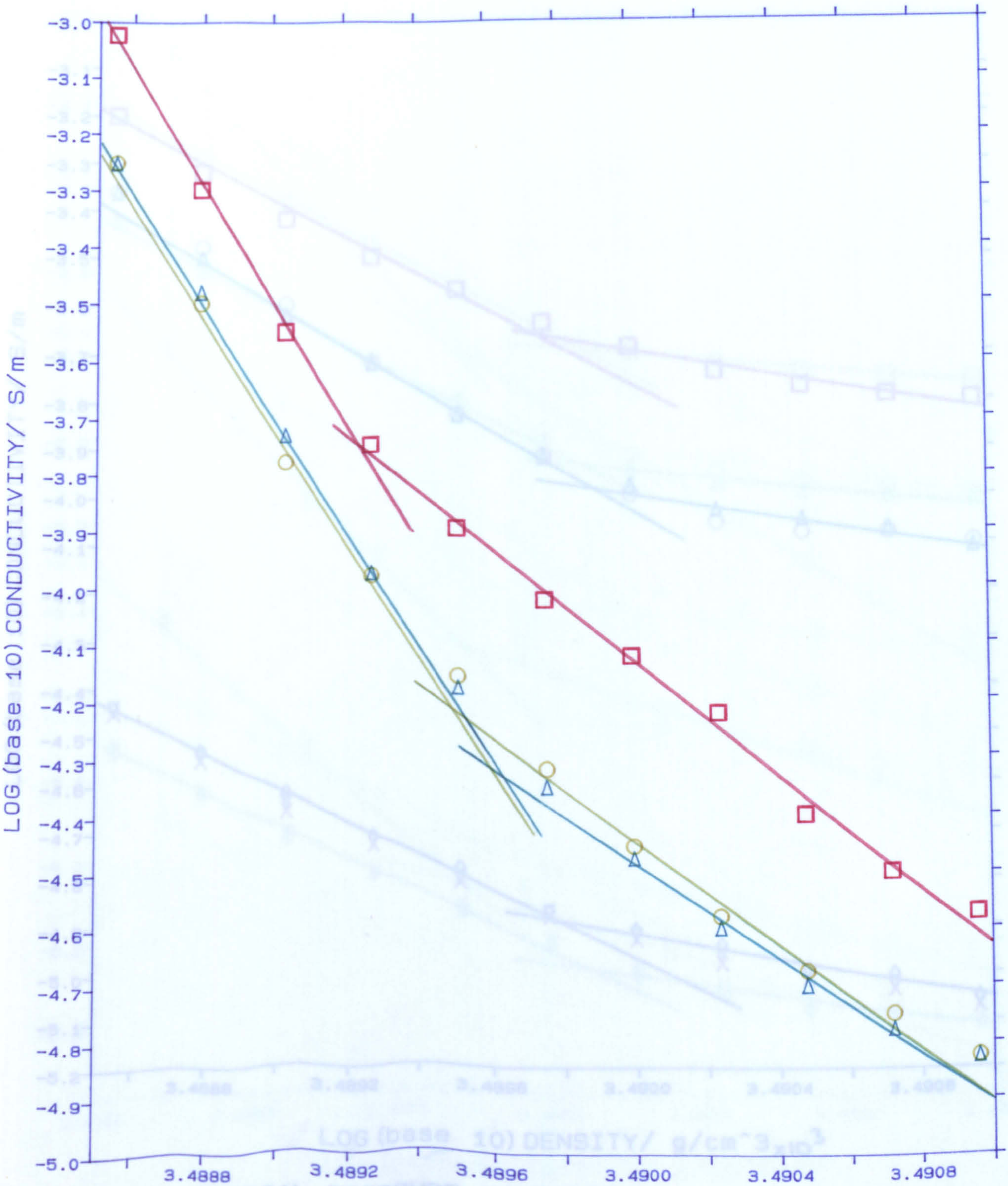
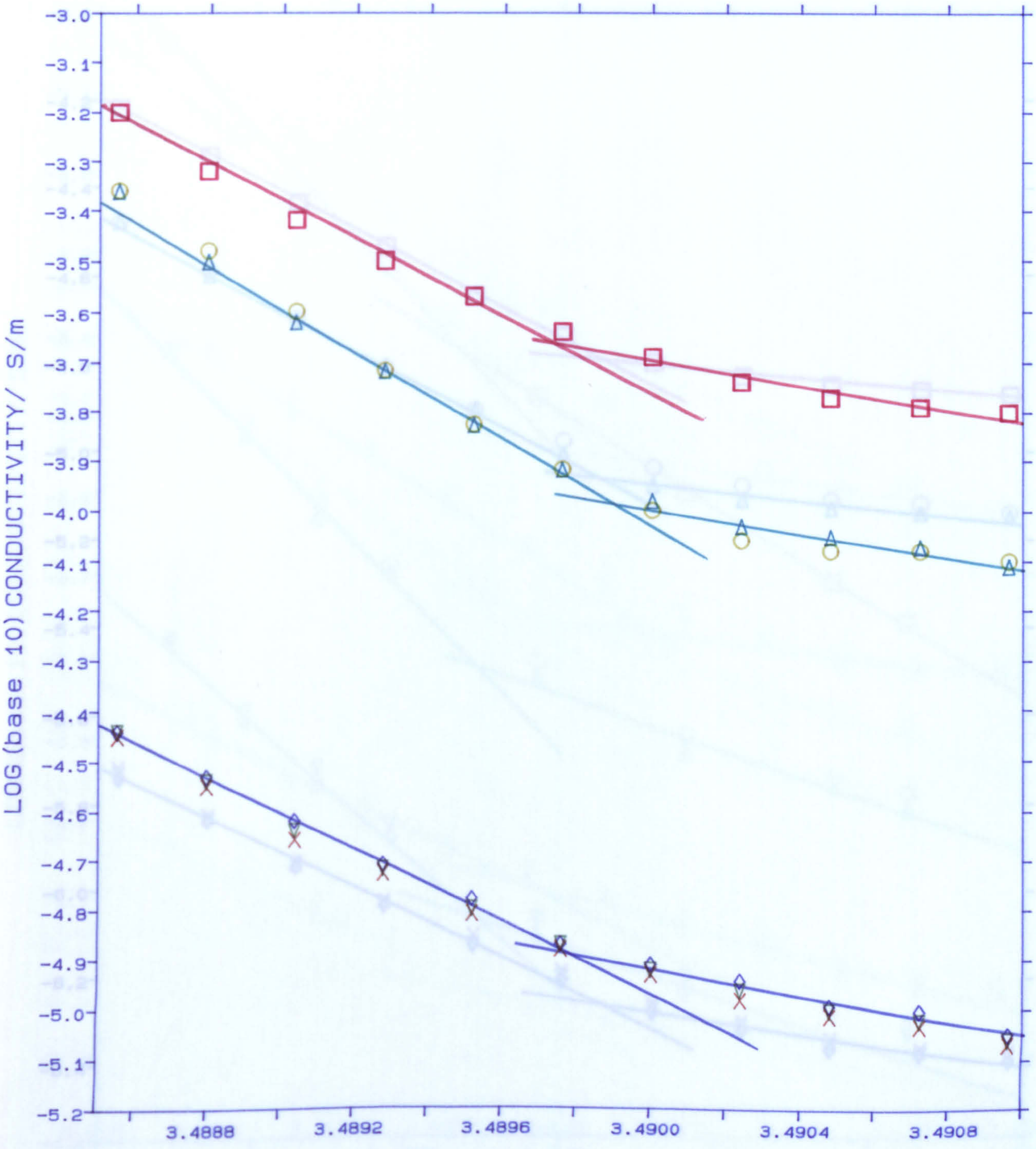
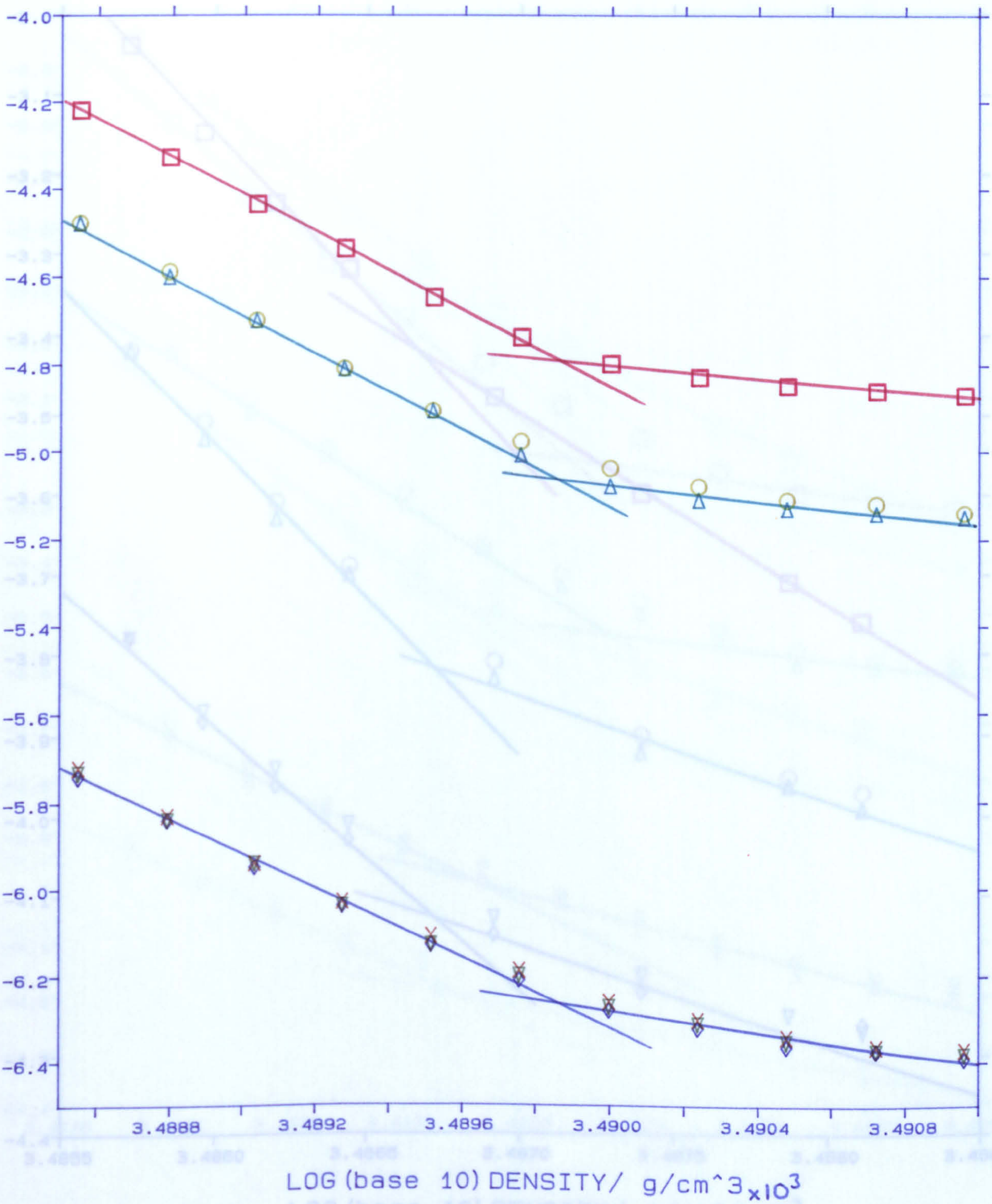


FIG. F9.8. MER21DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE



- □ INITIAL PRESSURE VARIATION
- ○ POST-PRESSURE TRANSIENT VARIATION
- △ POST-PRESSURE TRANSIENT VARIATION
- ▽ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
- ◇ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
- X POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F9.9. MER22DEN CONDUCTIVITY/DENSITY VARIATION FOR DISTILLED WATER SATURATED ORTHOPYROXENITE



□ INITIAL PRESSURE VARIATION
○ POST-PRESSURE TRANSIENT VARIATION
△ POST-PRESSURE TRANSIENT VARIATION
▽ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
◇ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
X POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F9.10. A8BDEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

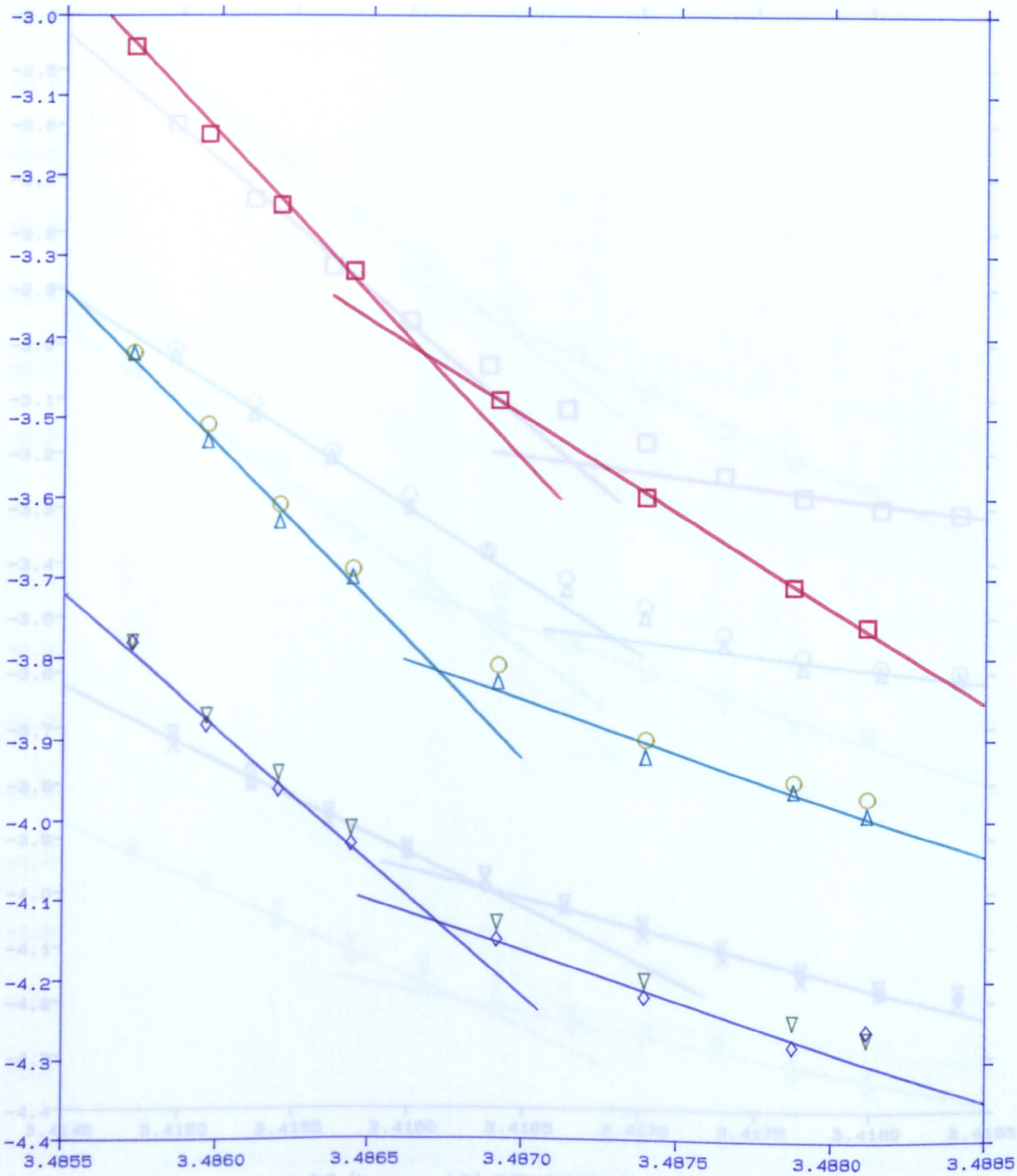
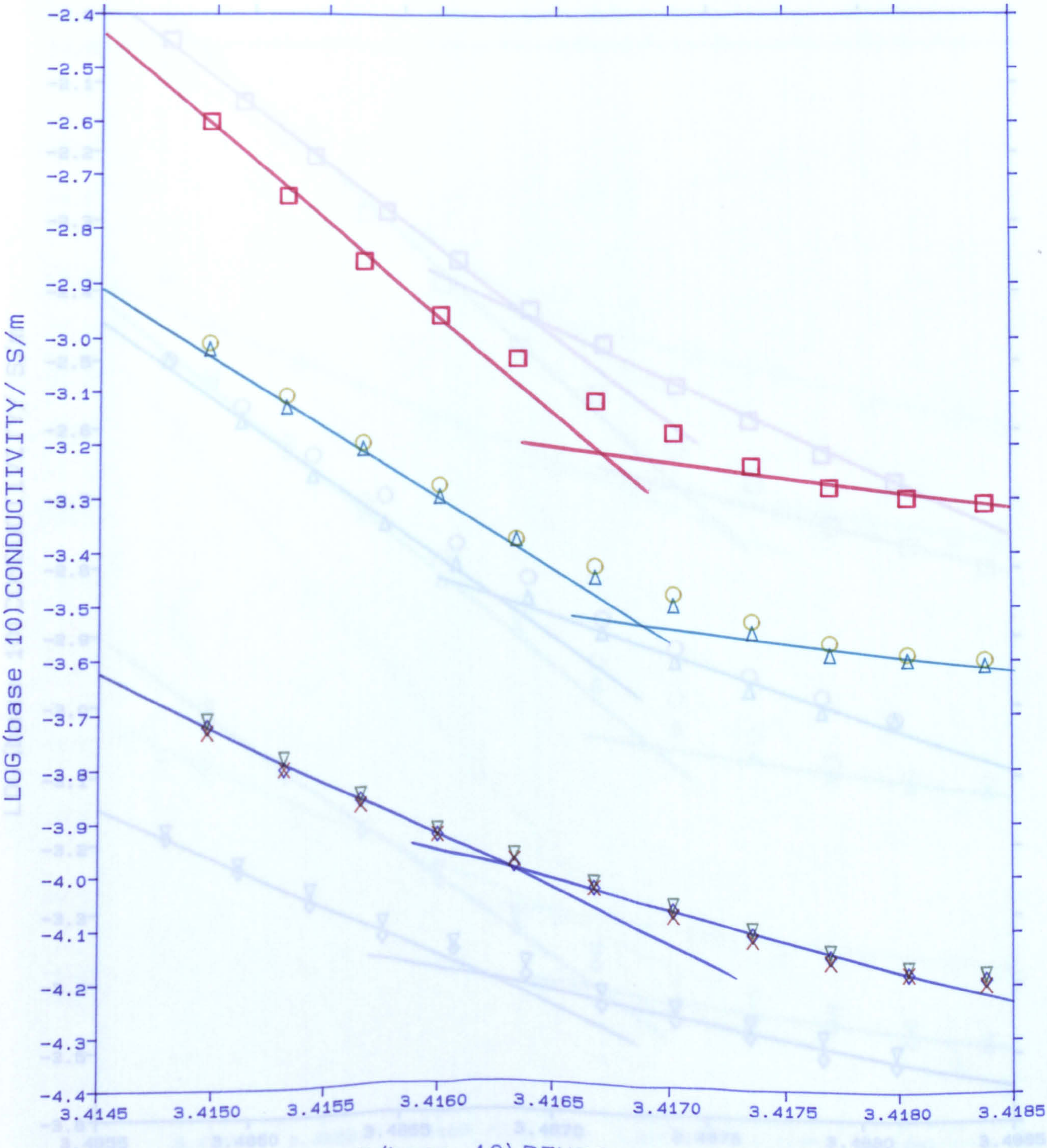


FIG. F9.11. L33BDEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE



- □ INITIAL PRESSURE VARIATION
- ○ POST-PRESSURE TRANSIENT VARIATION
- △ POST-PRESSURE TRANSIENT VARIATION
- ▽ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
- ◇ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
- X POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F9.12. L32DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

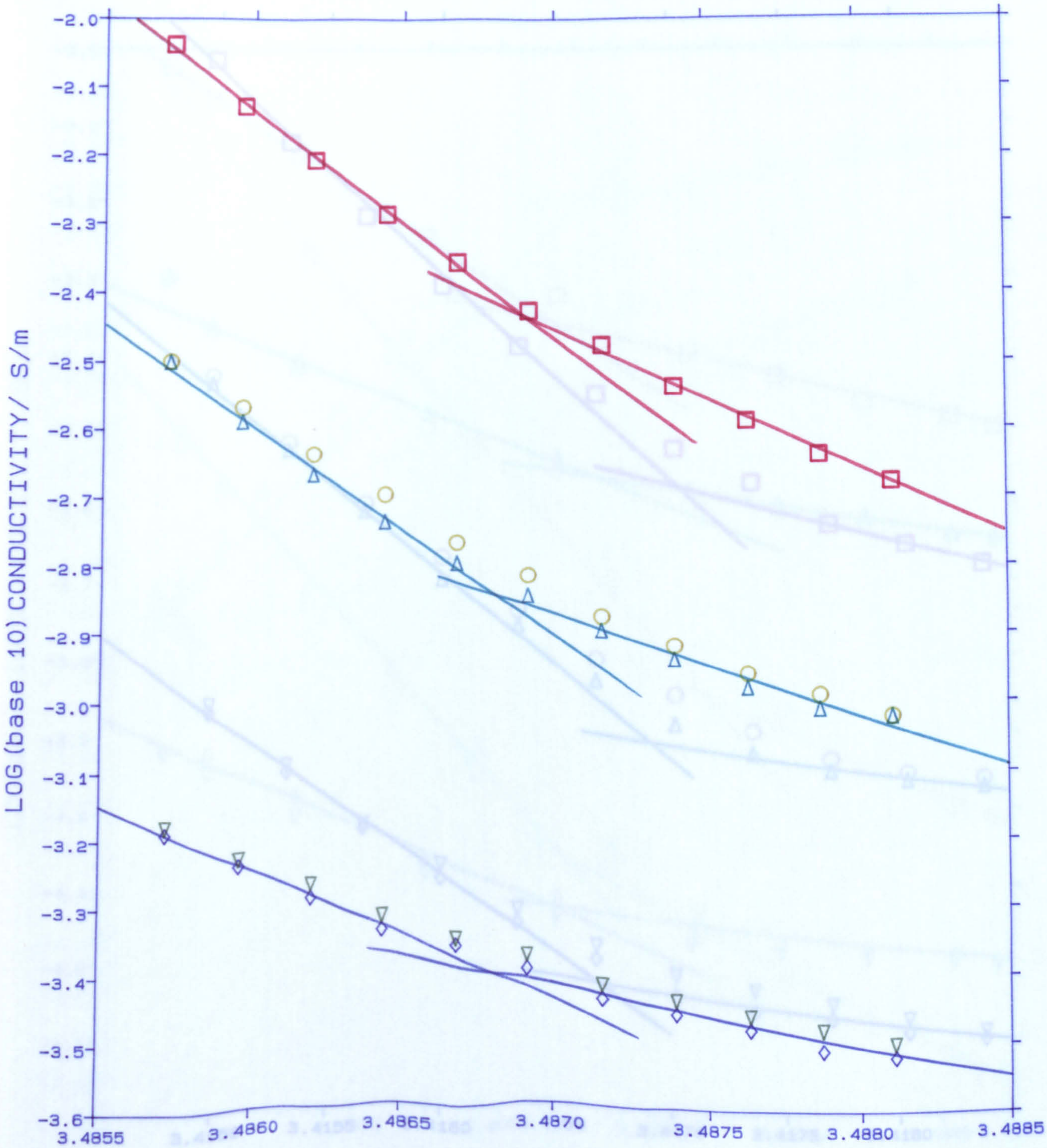
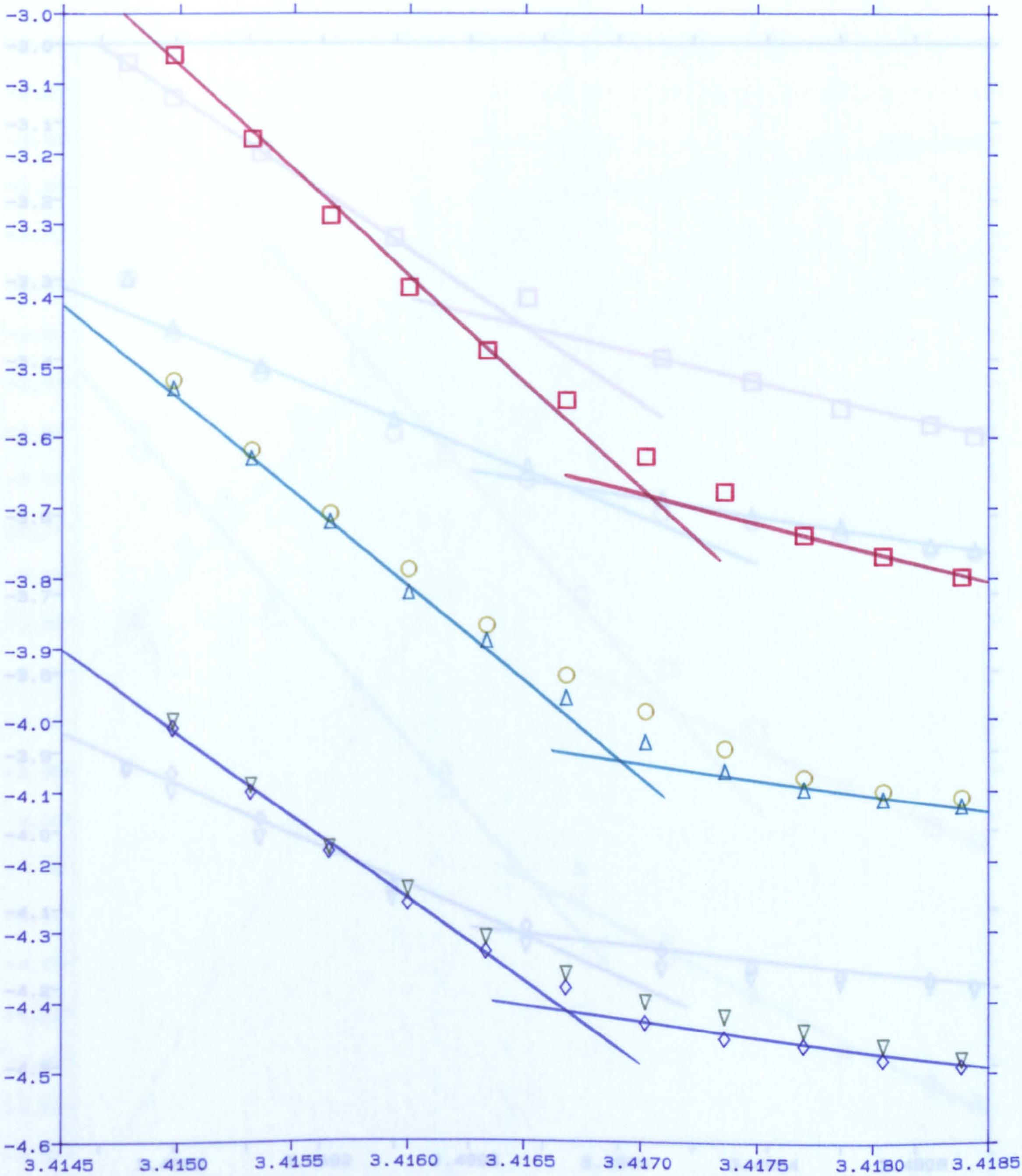
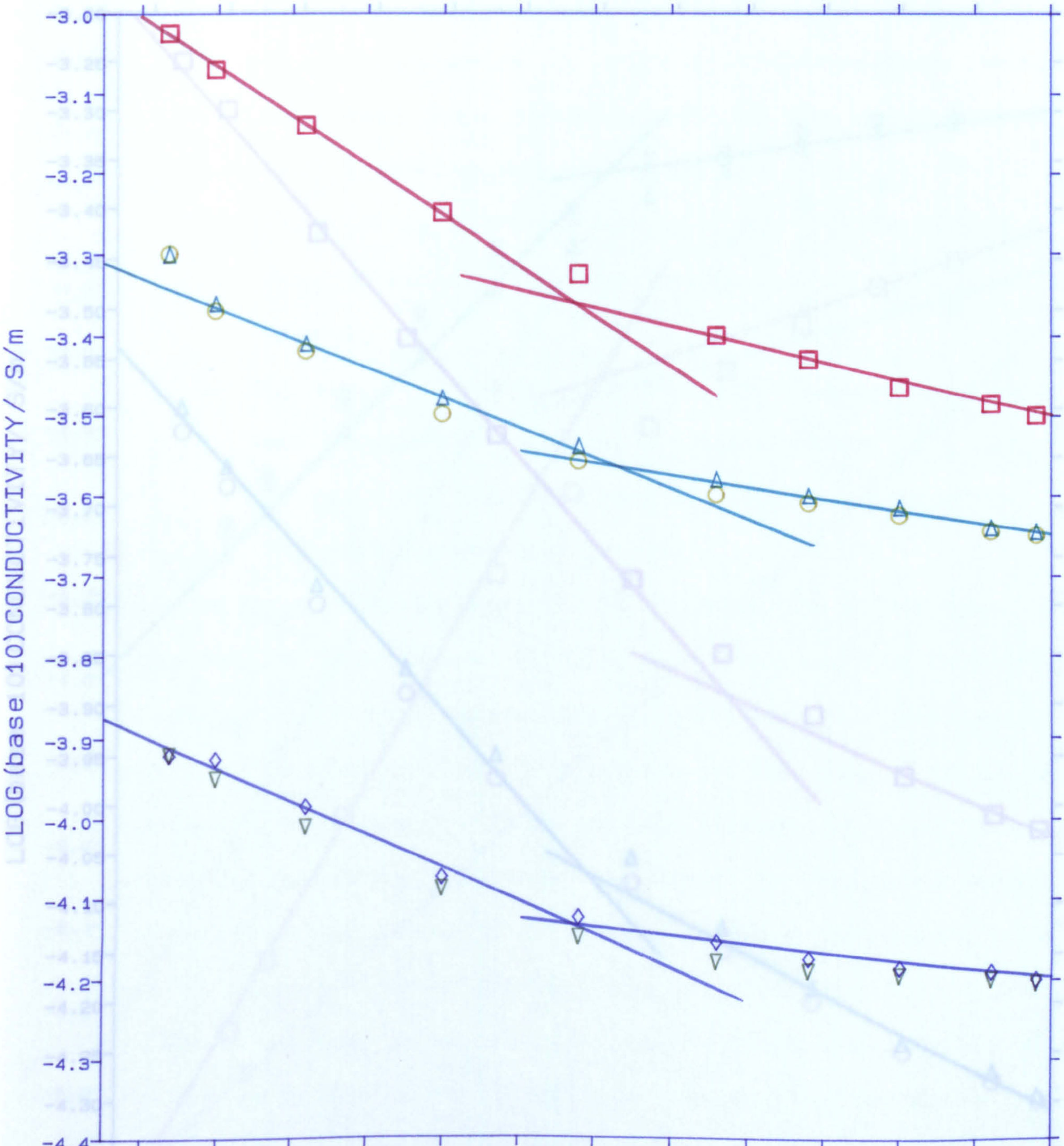


FIG. F9.13. V1DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRANITE



□ INITIAL PRESSURE VARIATION
○ POST-PRESSURE TRANSIENT VARIATION
△ POST-PRESSURE TRANSIENT VARIATION
▽ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
◇ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F9.14. GGRAN1DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M SOLUTION SATURATED GRAPHITIC GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATIONS



□ INITIAL PRESSURE VARIATION
○ POST-PRESSURE TRANSIENT VARIATION
△ POST-PRESSURE TRANSIENT VARIATION
▽ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION
◇ POST-PRESSURE AND TEMPERATURE TRANSIENT VARIATION

FIG. F9.15. GGRAN2DEN CONDUCTIVITY/DENSITY VARIATION FOR 0.5M NaCl SOLUTION SATURATED GRAPHITIC GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATION

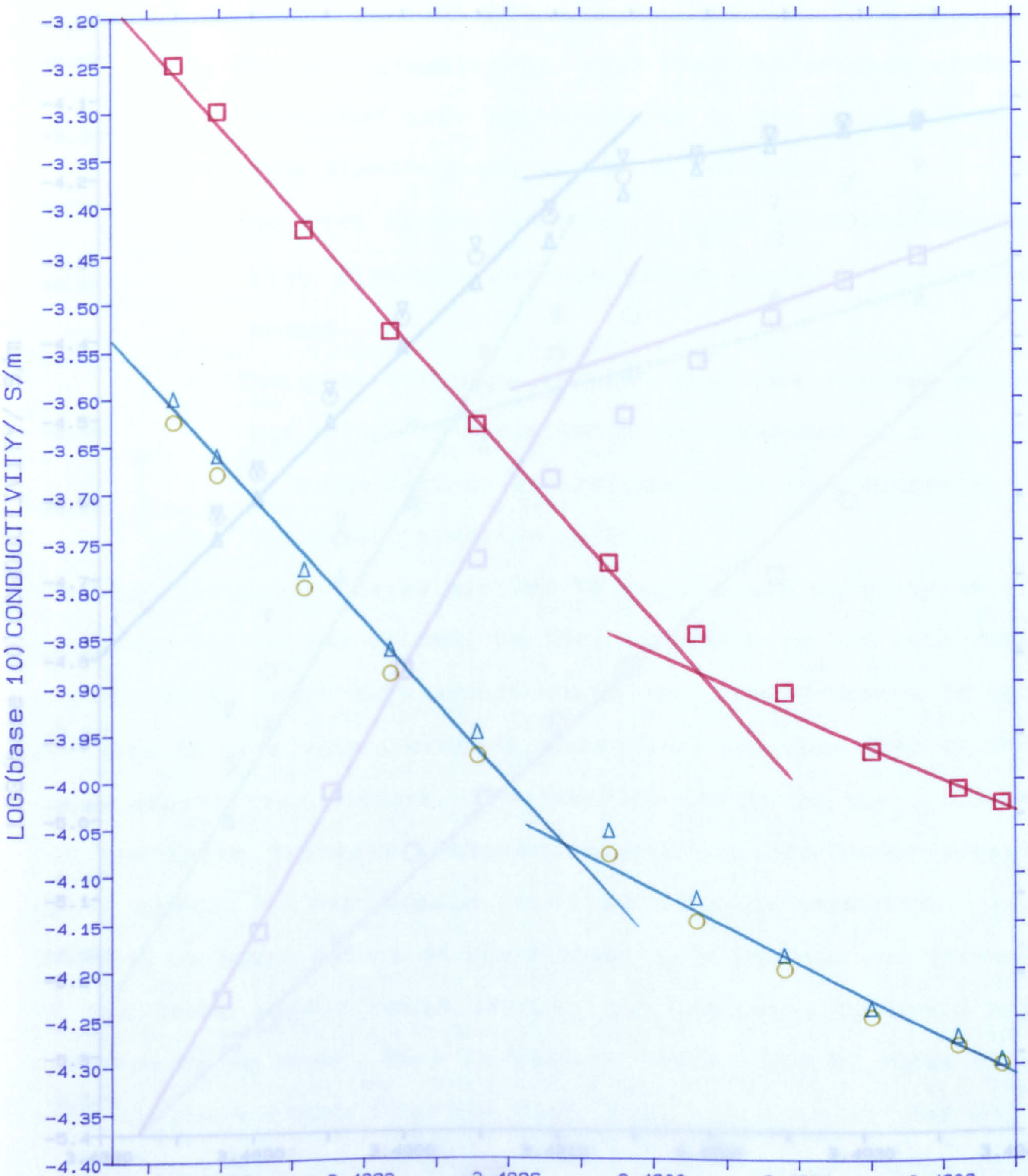


FIG. F9.16. GGRAN3DEN CONDUCTIVITY/DENSITY VARIATION OF DRY GRAPHITIC GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATIONS

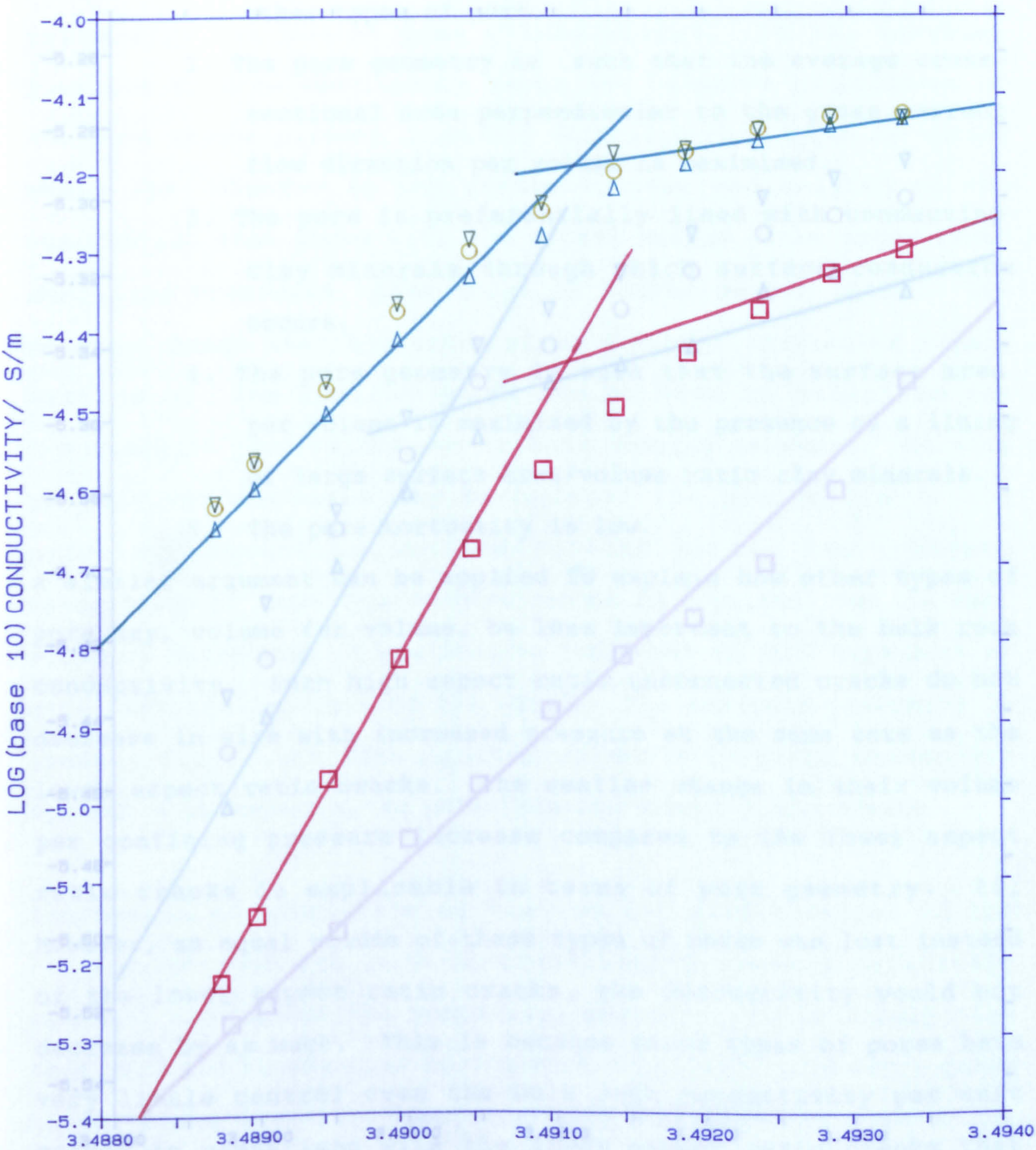
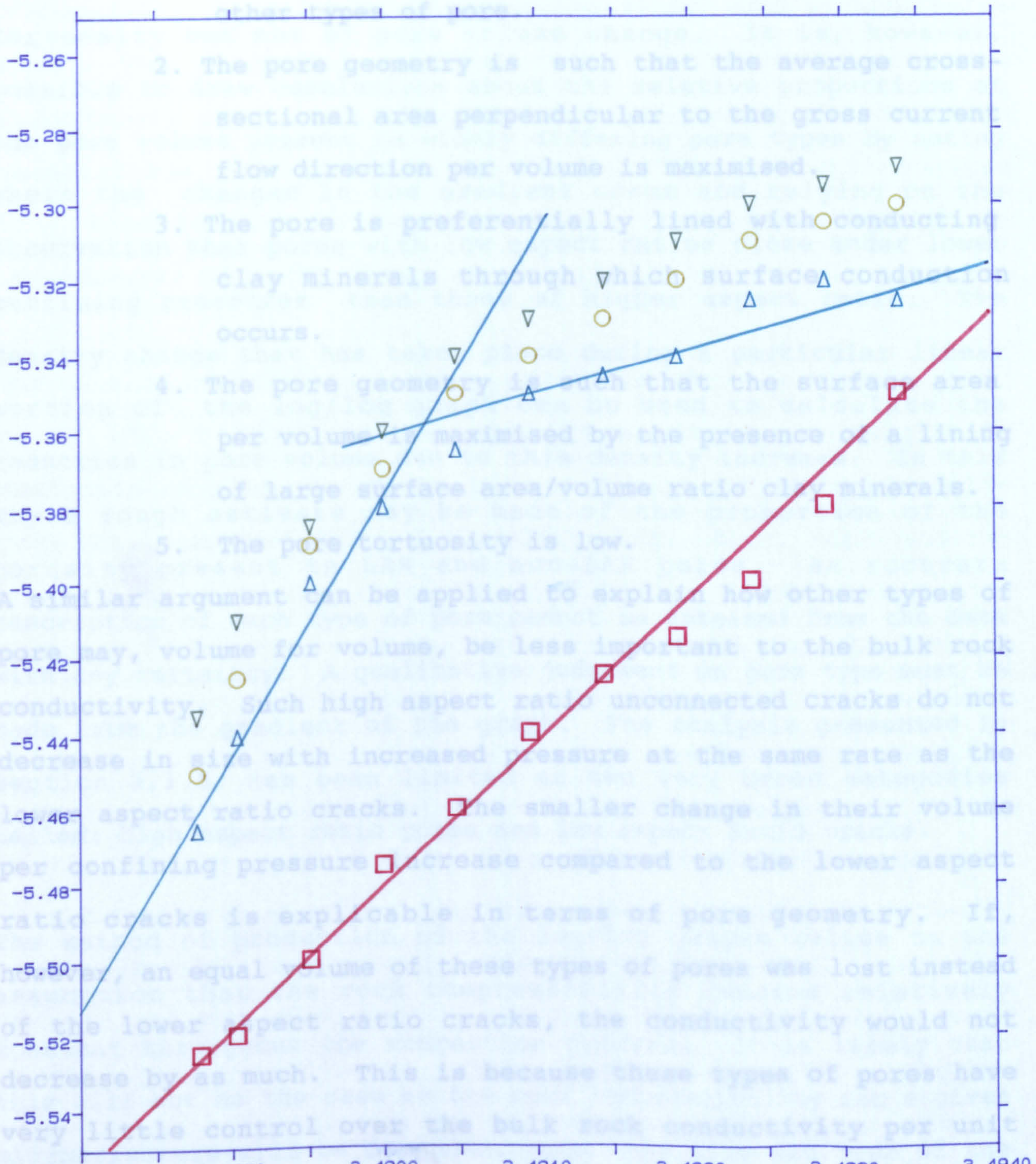


FIG. F9.17. GGRAN4DEN CONDUCTIVITY/DENSITY VARIATION OF DRY GRAPHITIC GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATIONS



1. The pore topology is such that it enables pores to be well connected to other pores of the same type or to act as good 'communicating' pores between other types of pore.
2. The pore geometry is such that the average cross-sectional area perpendicular to the gross current flow direction per volume is maximised.
3. The pore is preferentially lined with conducting clay minerals through which surface conduction occurs.
4. The pore geometry is such that the surface area per volume is maximised by the presence of a lining of large surface area/volume ratio clay minerals.
5. The pore tortuosity is low.

A similar argument can be applied to explain how other types of pore may, volume for volume, be less important to the bulk rock conductivity. Such high aspect ratio unconnected cracks do not decrease in size with increased pressure at the same rate as the lower aspect ratio cracks. The smaller change in their volume per confining pressure increase compared to the lower aspect ratio cracks is explicable in terms of pore geometry. If, however, an equal volume of these types of pores was lost instead of the lower aspect ratio cracks, the conductivity would not decrease by as much. This is because these types of pores have very little control over the bulk rock conductivity per unit volume in comparison with the lower aspect ratio cracks that fulfil some of the five points listed above. The gradient of the log/log graph will be lower as the decrease in conductivity per unit pore volume lost, and therefore per unit density increase, is less for such pore geometries.

It is important to note that the gradient of the log/log graph is diagnostic of pore geometry, topology, connectivity and tortuosity but not of pore volume change. It is, however, possible to draw conclusions about the relative proportions of the pore volume present in widely differing pore types by noting where the changes in the gradient occur and relying on the observation that pores with low aspect ratios close under lower confining pressures than those of higher aspect ratio. The density change that has taken place during a particular linear portion of the log/log graph can be used to calculate the reduction in pore volume due to this density increase. In this way a rough estimate may be made of the proportion of the porosity present in LAR and non-LAR pores. An accurate description of each type of pore cannot be obtained from the data with any certainty. A qualitative judgement on pore type must be made from the gradient of the graph. The analysis presented in section 9.1.5. has been limited to two very broad categories called; high aspect ratio pores and low aspect ratio cracks.

The method of production of the log/log graphs relies on the assumption that the rock compressibility remains relatively constant throughout the compaction process. It is likely that this will not be the case as the rock compressibility for a given microstructure will be dependent upon the size and type of the remaining porosity. Ideally the variation of the rock compressibility with pressure and temperature would be known and applied to producing log/log variations of conductivity against density that do not rely on the assumption of a constant

compressibility. The rock compressibility as a function of pressure and temperature can be calculated from values of V_p and V_s. The 'simultaneous' measurement of rock electrical conductivity, V_p and V_s would be necessary for such a thing to be done. The practical considerations of inserting acoustic transducers into a cell designed for high pressures and high temperatures such as those reached in this work will require considerable work. Some work has already been done up to temperatures of 250°C (**Evans et al., 1982**).

The density transformed log/log graphs are shown as figures F9.4. to F9.17. Straight line portions exist in the graphs of metabasic samples of amphibolite, granulite and orthopyroxenite from the Ivrea Zone IVD1, IVD2, IVE1, IVE2, MER21, MER22 and the samples of graphite garnet granulite GGRAN1, GGRAN2, GGRAN3 and GGRAN4. Linear portions may exist in the other graphs but are masked by a gradual transition between them. This is probably the case of the amphibolite gneiss samples A8B and L32. Alternatively a wide range of pore type may be present in the rock. If there is a wide distribution of different pore geometries with no dominating types then it would be expected that a smoothly varying log(conductivity)/log(density) be found. This is the variation seen in the case of the two acidic rocks studied (the granodiorite, L33B and the granite V1). A graph with two well delineated linear portions indicates a bimodal distribution of pore type, where type refers to those pores that control the rock conductivity and those that do not have that amount of control. The results seem to indicate that bimodal pore type distributions are more likely to be encountered in the case of metabasic rocks than the acidic rocks.

All of the graphs named above have two linear portions. The first has a greater negative gradient than the second in all cases. This behaviour can be qualitatively described in the following way: The first linear portion refers to the situation in which the rock experiences a large fractional decrease in the electrical conductivity per unit pore volume decrease because the pores that are closing have great influence upon the bulk rock conductivity. These pores are most likely to be of low aspect ratio, having a thin, dendritic, crack-like character. The crack-like pores would be more likely to be highly interconnected and although complex such a connectivity would provide a relatively low tortuosity. Low aspect ratio cracks such as these close at lower confining pressures than those of a more spherical nature (**Walsh, 1965**). The higher negative gradient at lower pressures than that at higher pressures in the log/log graphs is therefore consistent with Walsh's hypothesis and tends to increase the confidence in the log density transformation as a way of displaying confining pressure data.

The second linear portion of the graphs (occurring at higher confining pressures) has a smaller negative gradient as has already been pointed out. It is in this region that the fractional decrease on the electrical conductivity per unit increase in density (unit decrease in pore volume) is less than before. The pores that close during this behaviour do not have as much control over the bulk rock electrical conductivity as those which were closing at lower pressures. The pore voids are likely to have higher aspect ratios, smaller surface area to

volume ratios, low inter-connectivity and greater sphericity. They are also more likely to survive to relatively higher pressures due to :

1. Walsh's theory
2. Badly or unconnected fluid filled pores, when subjected to a high confining pressure external to the rock that they occupy, produce an increase in the pore pressure in the isolated pore fluid which acts to oppose the closure of the pore by the confining pressure. This can be likened to trying to compress fluid in a syringe the port of which is tightly closed. The same argument applies to a gas saturated pore but in this case the reaction of the externally pressured pore gas against closure is less efficient due to the greater compressibility of the gas. This can also be demonstrated using a gas or gas and liquid filled syringe. The ease of movement of the syringe plunger is proportional to the fraction of gas to fluid for any given uncompressed volume of gas/liquid mixture.

The delineation of two linear portions in the log/log graphs indicates a bimodal pore geometry distribution, within the range of pores closeable up to 0.2 GPa. The samples showing a steady decrease in gradient remain consistent with Walsh's hypothesis. The lack of well delineated linear portions indicates a more comprehensive distribution of pore geometries with a smoothly varying transition from one geometry to another.

Figures F9.16. and F9.17. show increases in conductivity due to the presence of graphite. This effect will be described later in this section. In these two cases the decrease in positive gradient with increase in density is consisitent with Walsh's hypothesis.

The staged decrease in conductivity in basic rocks compared to acidic rocks was noted in many of the samples studied by **Jackson (1984)**. He attributed the staged decrease and the overall smaller basic rock conductivity decrease with pressure to the presence of clay mineralogy lining the thin cracks retaining the electrical patency of the crack when it had been physically closed. This work has shown that;

1. The overall initial conductivity for the metabasic rocks is higher than that for the acidic rocks despite similar porosities (Table T9.1., parameter a) indicating that there needs to be an additional conduction mechanism present in these.
2. The rate of decrease of conductivity with pressure for all types of rock is similar although there is a small tendency for the acidic rocks to decrease faster than the metabasic rocks especially for the post-temperature and post-pressure transient variations (Table T9.1., b parameter). The small difference can be explained in terms of the acidic rocks having a greater proportion of low aspect cracks (Table T9.3.) that are more easily closed for any given confining pressure. The reduction in conductivity decrease due to clay mineralogies keeping physically closed

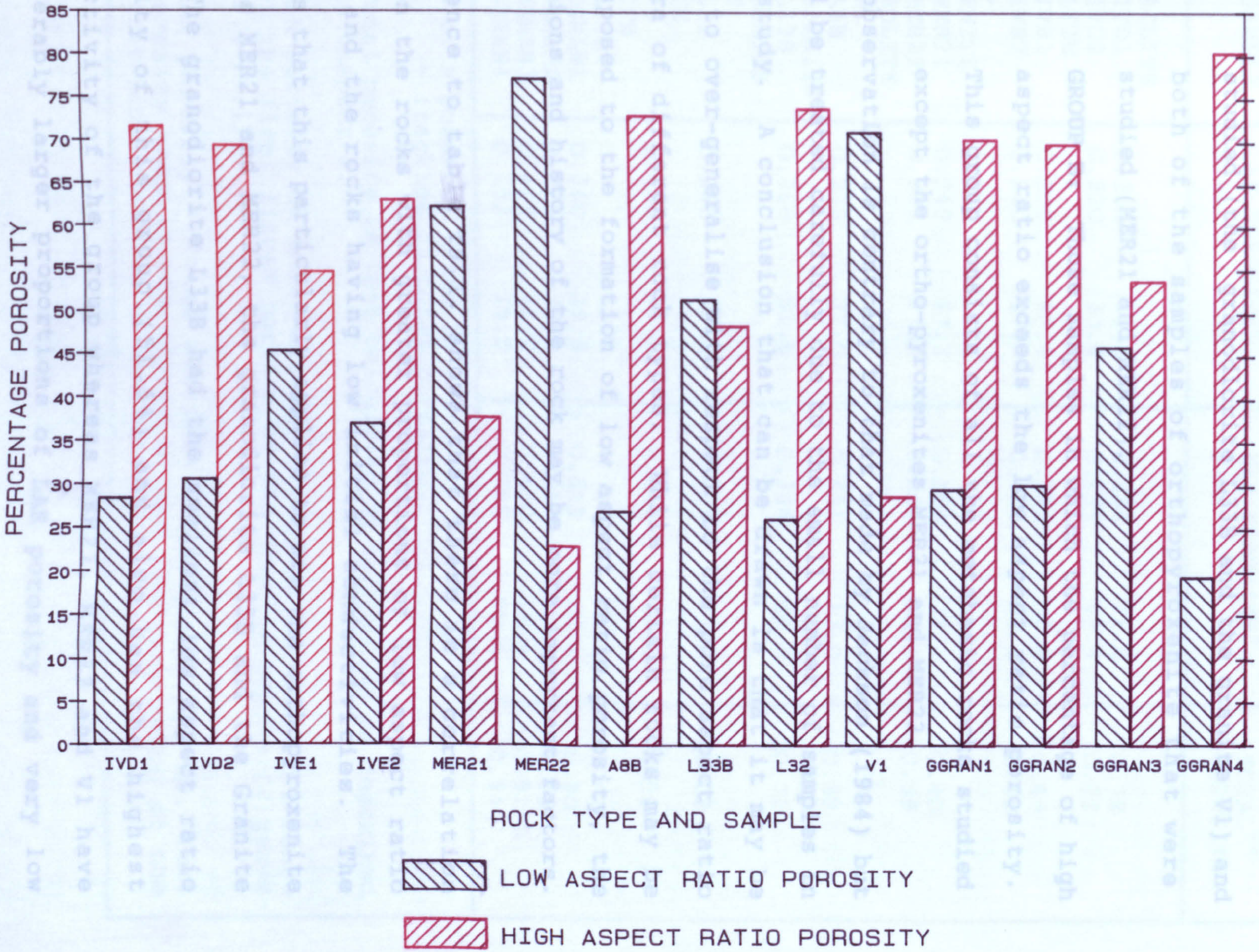
pores electrically open, therefore if present, is not so large as to be considered to be a major effect, as envisaged by **Jackson (1984)**, however the presence of some enhanced conduction is very significant and remains relatively unaffected by the increase in confining pressure.

9.1.5. ESTIMATIONS OF CRACK ASPECT RATIO PROPORTIONS

Estimations of the porosities due to high and low aspect ratio pore volumes are given in table T9.3. The figures have been calculated from knowledge of the initial density of the rock, the density of the pore fluid and the apparent reduction in rock density due to application of a confining pressure.

The high aspect ratio region has been taken as that up to an obvious 'knee' in those graphs showing bimodal dual linear regions and has been taken as that up to the extrapolated intercept of two linear fits to the low and high pressure curves in the case of those curves showing no obvious 'knee'. The estimate in the latter case will necessarily be less accurate. The accuracy of the estimation technique has been calculated to have an error of less than 5% in all cases, more accurate in some cases. It will be noticed that in some cases there is a migration in the position of the 'knee' after the pressure transient and again after the temperature transient. The migration occurs in the direction of lower density. This suggests that post-transient rock samples have had some of their low aspect ratio cracks permanently closed by their high temperature and/or pressure history. Figure F9.18. shows a

FIGURE F9.18. RELATIVE OCCURRENCES FOR DIFFERENT TYPES OF POROSITY



graphical representation of table T9.3.

The samples can be divided into two groups;

GROUP A: Those samples in which the percentage of low aspect ratio porosity exceeds that of high aspect ratio. This group consists of both of the acidic rocks studied (the granodiorite L33B and the granite V1) and both of the samples of orthopyroxenite that were studied (MER21 and MER22).

GROUP B: Those samples in which the percentage of high aspect ratio exceeds the low aspect ratio porosity. This group consists of all the metabasic rocks studied except the ortho-pyroxenites MER21 and MER22.

This observation is contrary to that made by **Jackson (1984)** but should be treated carefully due to the small number of samples in this study. A conclusion that can be drawn is that it may be false to over-generalise with respect to the pore aspect ratio spectra of different rock types. While certain rocks may be predisposed to the formation of low aspect ratio porosity, the conditions and history of the rock may be more important factors.

Reference to table T9.1. shows that there is a correlation between the rocks with greater proportions of low aspect ratio pores and the rocks having low initial conductivities. The samples that this particularly applies to are the orthopyroxenite samples MER21 and MER22, the granodiorite L33B and the Granite V1. The granodiorite L33B had the smallest low aspect ratio porosity of this group (51.5%) and also has the highest conductivity of the group whereas MER21, MER22 and V1 have considerably larger proportions of LAR porosity and very low

initial conductivities. There seems to be no correlation between the proportion of aspect ratios and the rate of reduction of conductivity with pressure (parameter b of table T9.1.).

TABLE T9.3. POROSITIES CALCULATED FROM LOG/LOG GRAPHS

SAMPLE	LAR POROSITY[+] p.u.	%	HAR POROSITY[*] p.u.	%	TOTAL[++] POROSITY
AMPHIBOLITE					
IVD1	0.221	28.4	0.558	71.6	0.78
IVD2	0.220	30.6	0.499	69.4	0.72
GRANULITE					
IVE1	0.308	45.4	0.372	54.6	0.68
IVE2	0.240	37.0	0.409	63.0	0.65
ORTHO-PYROXENITE					
MER21	0.249	62.3	0.150	37.7	0.40
MER22	0.301	77.2	0.089	22.8	0.39
AMPHIBOLITE GNEISS					
A8B	0.238	27.0	0.651	73.0	0.89
GRANODIORITE					
L33B	0.366	51.5	0.344	48.5	0.71
AMPHIBOLITE GNEISS					
L32	0.266	26.1	0.754	73.9	1.02
GRANITE					
V1	0.355	71.2	0.144	28.8	0.50
GRAPHITE GRANULITE					
GGRAN1	0.260	29.6	0.619	70.4	0.88
GGRAN2	0.279	30.1	0.651	69.9	0.93
GGRAN3	0.419	46.1	0.490	53.9	0.91
GGRAN4	0.183	19.5	0.757	80.5	0.94

p.u. = porosity units in percent
% = fraction of total porosity

++ Total porosity is that calculated by saturation as described in chapter 4.

+ LAR porosity is that porosity closed by the pressure range of the first linear portion of the density graphs (Figures F9.4-F9.17).

* HAR porosity = Total Porosity - LAR porosity, ie. that porosity that is not closed by the pressure range of the first linear portion of the density graphs.

9.1.6. MORE EVIDENCE FOR ENHANCED CONDUCTION IN METABASIC ROCKS

Many of the rock samples were of a type that would be expected to exhibit an enhanced conductivity (Jackson, 1984; Lee, 1977). The conductivities of these more basic rocks have been shown to;

- not obey Archie's Law in a similar manner to that observed by Lee (1977).
- have measured values of conductivity significantly greater than that predicted by the use of Archie's law alone. Such an observation is a strong indication for the presence of conduction mechanisms other than electrolyte conduction.
- not show a decrease with pressure as great as most acidic rocks. Previous works have pointed out that this is despite the demonstration that the basic rocks contain a greater proportion of low aspect ratio cracks which would have been more likely to cause a greater variation of the basic rock conductivity with pressure, (Jackson, 1984). This observation has been used as evidence for a basic rock enhanced conduction mechanism.

The results of this work (presented in section 9.1.5.) do not agree with this last point. Reference to figure F9.18. shows that the more basic rocks have proportionably greater amounts of HAR pores than LAR cracks than the more acidic rocks. There were, however, only two acidic rocks included in this work so the comparison population is too small for any solid inferences to be made. It may be the case that the recent rock history

controls the predominance of low aspect ratio fracture porosity. If this is the case then a large number of samples would be needed to distinguish those types of rocks more predisposed to the formation of low aspect ratio pores.

All the measured conductivities were compared with the conductivities predicted from porosity and pore fluid conductivity data using Archie's Law.

$$\sigma_r = \sigma_p \ \phi^a \qquad \text{E9.13.}$$

where σ_r is the predicted conductivity of the saturated sample,

σ_p is the pore fluid conductivity and

ϕ is the rock porosity.

a is a constant. The results of observed and predicted conductivities are shown together in table T9.4.

The most obvious conclusion to be drawn from the comparison between the measured and predicted conductivities is that the measured conductivity is much greater than can be explained by electrolyte conduction alone in the basic rocks. The measured values of conductivity in the basic granulite, IVD1 and basic amphibolite, IVE1, are 9.4 and 25.9 times greater than the predicted value using an exponent of 2. The possibility that the values do not agree due to the assumption that a value of a=2 being false in rocks with a significant proportion of LAR cracks as well as HAR pores has been investigated below. The apparent value for "a" can be calculated from the measured conductivity of the rock and brine, and the measured rock porosity. This shows values for "a" consistently below 2 as would be expected. Recalculation using the HAR conductivity and associated porosity

from the pressure variations shows the values of "a" to be improved and in some cases approach a=2. The observation that the corrected values of "a" become closer to 2 but still fall short of a=2 in most cases by 0.2 to 0.45 demonstrates that although the assumption that a=2 applies in all rocks is false it only accounts for some of the shortfall in the predicted conductivity. This data is thus a good indication that the higher observed conductivities are in large part due to some conduction mechanism other than electrolytic brine pore fluid conduction.

A value of "a" equal to two was found to be a good fit for most rocks with spherical pores (e.g. the graphite granulite samples GGRAN1 and GGRAN2)(Brace, Orange and Madden, 1965) but the value of "a" decreases in the samples with a pore aspect ratio spectrum biassed towards low aspect ratio cracks (e.g. the orthopyroxenite, MER21; the basic granulite, IVE1; the granodiorite L33B and the granite, V1). The results are shown in table T9.4.

TABLE T9.4. ARCHIE EQUATION THEORETICAL PREDICTIONS COMPARED WITH MEASURED VALUES OF CONDUCTIVITY

SAMPLE	MEASURED CONDUCTIVITY $S\ m^{-1}$ $X10^{-3}$	PREDICTED CONDUCTIVITY a=2, TOTAL POROSITY $S\ m^{-1}$ $X10^{-3}$	GENERATED a ON TOTAL POROSITY	GENERATED a ON HAR POROSITY
BASIC GRANULITE				
IVD1	2.51	0.266	1.53	1.81
IVD2	0.32			
BASIC AMPHIBOLITE				
IVE1	5.25	0.202	1.34	1.58
IVE2	0.93			
ORTHO-PYROXENITE				
MER21	0.63	0.069	1.60	1.58
MER22	0.60	0.066	1.59	1.78
AMPHIBOLITE GNEISS				
A8B	0.91	0.345	1.79	1.87
GRANODIORITE				
L33B	2.51	0.220	1.51	1.58
AMPHIBOLITE GNEISS				
L32	8.91	0.454	1.35	1.48
GRANITE				
V1	0.87	0.109	1.61	1.49
GRAPHITE GRANULITE				
GGRAN1	0.95	0.338	1.78	1.83
GGRAN2	0.63	0.377	1.89	2.03

Equation E9.13. can be rewritten

$$\log_{10} \sigma_r = \log_{10} \sigma_p + a \log_{10} \phi \qquad \text{E9.14.}$$

Brace and Orange (1968) have suggested that the porosity should be split up into two components when applied to Archie's Law;

- the porosity due to 'cracks', low aspect ratio porosities responsible for the confining pressure variation of electrical conductivity at pressures *below 0.4 GPa.*

- the porosity due to 'pores', high aspect ratio porosities that are relatively unresponsive to pressure variations less than 0.4 GPa.

$$\sigma_r = \sigma_p \left\{ \phi_{LAR}^{b} + \phi_{HAR}^{a} \right\} \qquad \text{E9.15.}$$

The values (a and b) of the Archie exponents are different depending upon the predominating pore aspect ratio of the cracks and pores in the sample. It would therefore be expected that the value of these exponents would change from rock to rock and as a function of pressure. **Evans (1980)** used this theory to calculate a predicted Archie conductivity from a corrected value of porosity. The correction provided a better fit for the data in the case of high quartz rocks but worse for more basic rocks. This provides a further distinction between the electrical response of quartzose and relatively non-quartzose rocks and also points towards an enhanced conduction in more basic rocks. It can be seen from table T9.4. that the predicted value obtained from the use of Archie's equation with a=2 and the porosities obtained from the saturation method, produce values that underestimate the values obtained by measuring the rocks. This may be due to a combination of one or several of the following effects:

1. The value of the porosity being inaccurate. It has already been said in chapter 4 that the accuracy of low porosity measurements is probably around 10%. It is likely that the method will underestimate the porosity due to incomplete saturation of such tight rock samples. The underestimation error is multiplied by the value

of 'a'.

2. The assumption that a value of a=2 is inaccurate. The value of a=2 originally suggested by Archie was based upon experimental studies upon a specific range of frequencies, fluid types and conductivities, rock types and porosities concerned with water-bearing high porosity sedimentary rocks of inter est to the oil industry. The empirical equation with a=2 has, however, been very generally applied. It has been found to apply for a wider range of the above parameters in many cases such as some low porosity crystalline rocks (Brace et al., 1965; Brace and Orange, 1968) In other cases it has been applied without testing. There are, therefore, many cases where it doesn't apply and in such cases it is just the value of 'a' that needs to be changed before validity returns to the equation.

3. The rocks have an alternative conduction mechanism that is not modelled by the Archie equation or involving conduction through the bulk saturating fluid. Such an additional conduction mechanism may be responsible for some of the apparent extra measured conductivity.

4. Inaccurate values of the conductivity of the saturating fluid being used in predicting the conductivity. The solutions that were used to saturate the rocks were made up with standard *chemical techniques* using reagent quality, dry

chemicals and double distilled water. Corrections were made for buoyancy, dissolved gases, temperature effects and peri-mixing thermal effects. The conductivity was measured accurately using a calibrated dip-cell. The electrical conductivity was accurate to within 1%. It is possible, however, that the fluid saturating the rock sample could interact with the rock matrix to change the value of the electrical conductivity of the saturating fluid. Dissolution of minerals in the fluid may be able to raise the conductivity of the fluid sufficiently to make a considerable effect on the bulk conductivity, one which is not predicted by the Archie equation.

The discrepancy between the predicted and measured conductivities may not, therefore, be attributable solely to an extra conduction mechanism. This is demonstrated by the discrepancy present in the two acidic rocks studied. These are granite and granodiorite and unlikely to have any extra conduction mechanism operating as is demonstrated in section 9.3.

Table T9.4. also shows the values of 'a' that would be necessary for the Archie equation to hold. Two values are given. One to generate the conductivity from the total porosity. The other generates that conductivity judged to be due to only the non-LAR pores obtained by noting the value at a 'knee' in the log(conductivity) /log(density) graphs. It uses the correspondingly reduced porosity of non-LAR pores given in table T9.3. The total porosity values of 'a' are less than 2 as would be expected from the discrepancy between the a=2 predicted and

the measured values for the reasons given above. The suggestion that it is only the higher aspect ratio pores that a value of a=2 can be applied to has been tested by calculating the value of 'a' for reduced porosity. The values of 'a' are closer to 2 suggesting that Archie's law with a=2 is more applicable in these circumstances. The value could be still affected by:

1. Additional conduction mechanisms.
2. Inaccuracy in the calculation of the reduced porosity.
3. There being a considerable number of low aspect ratio cracks still present at the point where the reduced porosity and associated conductivity was taken.

Prediction of rock conductivity using Hoenig's theory (Hoenig, 1979; Evans et al, 1982) has not been possible due to the lack of complete seismic velocity data on the samples.

9.1.7. THE GRAPHITE GARNET GRANULITE SAMPLES

The graphs representing the graphite garnet granulite samples require special mention. The high proportion of graphite in these samples considerably affects the way in which they behave electrically. Reference to figures F8.11. and F8.12. shows that the overall magnitude of the electrical conductivity of these NaCl saturated samples remains in the range expected for a granulite type NaCl saturated rock. The variation with pressure does not behave so well. In the case of GGRAN1 (cut parallel with the graphite banding) the decrease in conductivity with increase in pressure is very much less than may be expected from this type of rock. It is also considerably less than the decrease

in conductivity with pressure of the sample GGRAN2 which was cut perpendicular to the planes of continuous graphite mineralogy. Experiments were done up to 0.4 GPa on two dry samples of the rock, GGRAN3 (cut parallel to the graphite planes) and GGRAN4 (cut perpendicular to the graphite planes). These variations are shown in figures F8.13. and F8.14. The most striking aspect of these graphs is that the conductivity increases with confining pressure rather than decreasing as is the case in the saturated rocks. The increase can be explained in terms of crack closure and with reference to the scanning electron micrographs discussed in section 4.6. The scanning electron micrographs show that the graphite forms along the surfaces of cracks that lie in a planar schistosity in the rock. The action of a confining pressure upon these cracks would be relatively rapid closure. Closure of the cracks would tend to bring more graphite into mutual contact and thus increase the conductivity by increasing the conductive pathways. This mechanism would tend to cause a greater variation of electrical conductivity with pressure on the sample with graphite planes lying in the direction of the electrical measurement than that where the planes of graphite are perpendicular to the direction of electrical measurement. Such a distinction can be seen by comparing the size of the variation in figures F8.13. and F8.14. These figures also show a pressure transient, but it is a transient **increase** as would be expected if grain slippage was bringing the graphite into better correspondence in an inelastic manner. The description of the manner of electrical conductivity variation with increasing confining pressure in dry graphite granulite is analogous to that of a saturated rock. This is interesting as it illustrates very

well that although the magnitude of the conductivity of rock samples depends upon the type and amount of highly conducting medium interspersed within the rock matrix, it is the properties of the rock matrix that defines the way that the conductivity of the rock varies with confining pressure. The measurements on dry graphite granulites are interesting in that they demonstrate the importance of anisotropy in the electrical conductivity of the rocks and the mechanism of pore closure resulting here in a conductivity increase rather than a decrease. The measurements are also important in respect of recent work by **Frost et al. (1989)**. Frost et al. have demonstrated that a thin layer of graphite can be deposited on the surface of rock grains under a wide range of conditions, not necessarily those associated with a stable graphite field. They mention that such thin surface films (about 1000 angstroms) if present on the surface of grains of diameter ranging from 0.1 - 10 mm should produce conductivities similar to those measured in the lower crustal high conductivity layers providing that the films remain unbroken. Indeed they speculate that such broken conduction paths would not be readily reconnected on laboratory raising of pressure. The proportion of graphite within the samples studied in this work is much greater and in this case it has been shown that reconnection can occur increasing the electrical conductivity as pressure is raised. The most important point arising from **Frost et al. (1989)** is that graphite may very well be more abundant in the lower crust than many have thought and this work has shown that such rocks as the graphite granulite studied here have an increasing conductivity with pressure in the dry state and as will be seen later has high temperature electrical conductivity values that can explain lower crustal high conductivity values. **Shankland (1989)** notes that

graphite could be a significant contributor to the overall electrical conductivity in both stable shield and active metamorphic zones while graphite has also been described by **Mathez and Delaney (1981)** and **Sternberg (1979)**.

9.2. THE EFFECT OF PORE FLUID PRESSURE UPON THE ELECTRICAL CONDUCTIVITY OF SATURATED ROCK SAMPLES AT RAISED CONFINING PRESSURES

Measurement of the response of the electrical conductivity of rock samples to changes in pore fluid pressures at low temperatures was not a primary aim of this work. Pore fluid pressure was necessary to ensure that the rock samples remained fully saturated at the high temperatures. The raising of pore fluid pressure was, therefore, necessary before the raising of temperature above 100^{o}C.

A comprehensive suite of pore fluid pressure measurements was not done due to the time that they would require. It was decided to make full measurements of the variation of electrical conductivity at high temperatures and confining pressures and at one, two or three pore fluid pressures. As experience was gained in making the experimental runs the number of temperature runs at different pore fluid pressures was increased from one to three. The effect of pore pressure changes in respect of the temperature variation of the rock electrical conductivity will be discussed in section 9.3. The variations of electrical conductivity with confining pressure at low temperatures were done at zero pore fluid pressure. The variations of electrical conductivity with confining pressure at raised temperatures were done at a constant pore fluid pressure of 0.08 GPa. The raising of pore fluid pressure before the temperature runs enabled the measurement of the variation of the saturated rock electrical conductivity on the initial raising of a pore fluid pressure at a constant confining pressure of 0.2 GPa. The variations are given as

figures F8.15. to F8.26. The figures show two apparently different types of behaviour, one a steady non-linear increase of electrical conductivity, the other is similar but has a large negative offset occurring at P_p = 0.1 GPa. These apparent differences are due to the way the data was collected. The post-offset variations on some of the figures are simply the post-temperature transient variations of the rock samples and can be explained by the order of the experiment.

A steady non-linear increase in electrical conductivity of the rock is observed in those samples that were taken to a confining pressure of 0.2 GPa and had their pore fluid pressure raised to 0.18 GPa (figures F8.15. to F8.18. and F8.20., samples IVD1, IVD2,IVE1,IVE2 and MER22).

The non-linear increase with offset is observed in those rock samples that were raised to a high confining pressure and then had their pore fluid pressure raised to 0.1 GPa. Temperature runs were then done with P_c = 0.2 GPa and P_p = 0.10 GPa. After the first rise in temperature the samples were left to undergo the temperature transient drop which remains whilst all the repeatable measurements for this set of confining pressure and pore fluid pressure were done. The sample was then raised in pore fluid pressure from 0.1 GPa to 0.18 GPa so that further temperature runs under the new conditions could be done. The variation in electrical conductivity during the increase of pore fluid pressure takes place from the post-transient state value for P_c = 0.2 GPa, P_p = 0.1 GPa, T = 25 ^oC. The offset seen in the pore fluid pressure variations is a relic of the temperature

transient drop. This type of variation can be seen in figures F8.19. and F8.21. to F8.26., samples MER21, A8B, L33B, L32, V1, GGRAN1 and GGRAN2. Samples GGRAN1 and GGRAN2 also had temperature variations subsequently measured at a pore fluid pressure of 0.08 GPa but the variation of electrical conductivity with pore fluid pressure while the pore fluid pressure was being reduced for these runs to be done was not measured. The undisturbed pore fluid pressure variation can be generated by adding the the transient drop electrical conductivity offset to the post-offset curve values.

It was found that electrical conductivity variations with temperature were affected by the magnitude of the pore fluid pressure. This effect was larger towards high temperatures and for the acidic rocks. The effect of pore fluid pressure upon the variation of the saturated rock electrical conductivitity in this context will be described in section 9.3. as part of the description on the electrical conductivity behaviour with changing temperature.

Comparison of figures F8.1. to F8.12. and F8.15. to F8.26. shows that the effect on the electrical conductivity of a given increment of pore pressure is not equal and opposite to the effect of an equal increment of confining pressure. This is equivalent to saying that the pore fluid pressure and confining pressures do not have equal and opposite effects upon the electrical conductiviiity of the saturated rocks. An effective pressure, P_e, can be defined as the pressure experienced by the rock sample as a result of the sum of the confining pressure and pore fluid pressure generated stresses. It can be stated thus;

$$P_e = P_c - a.P_p \qquad \text{E9.16.}$$

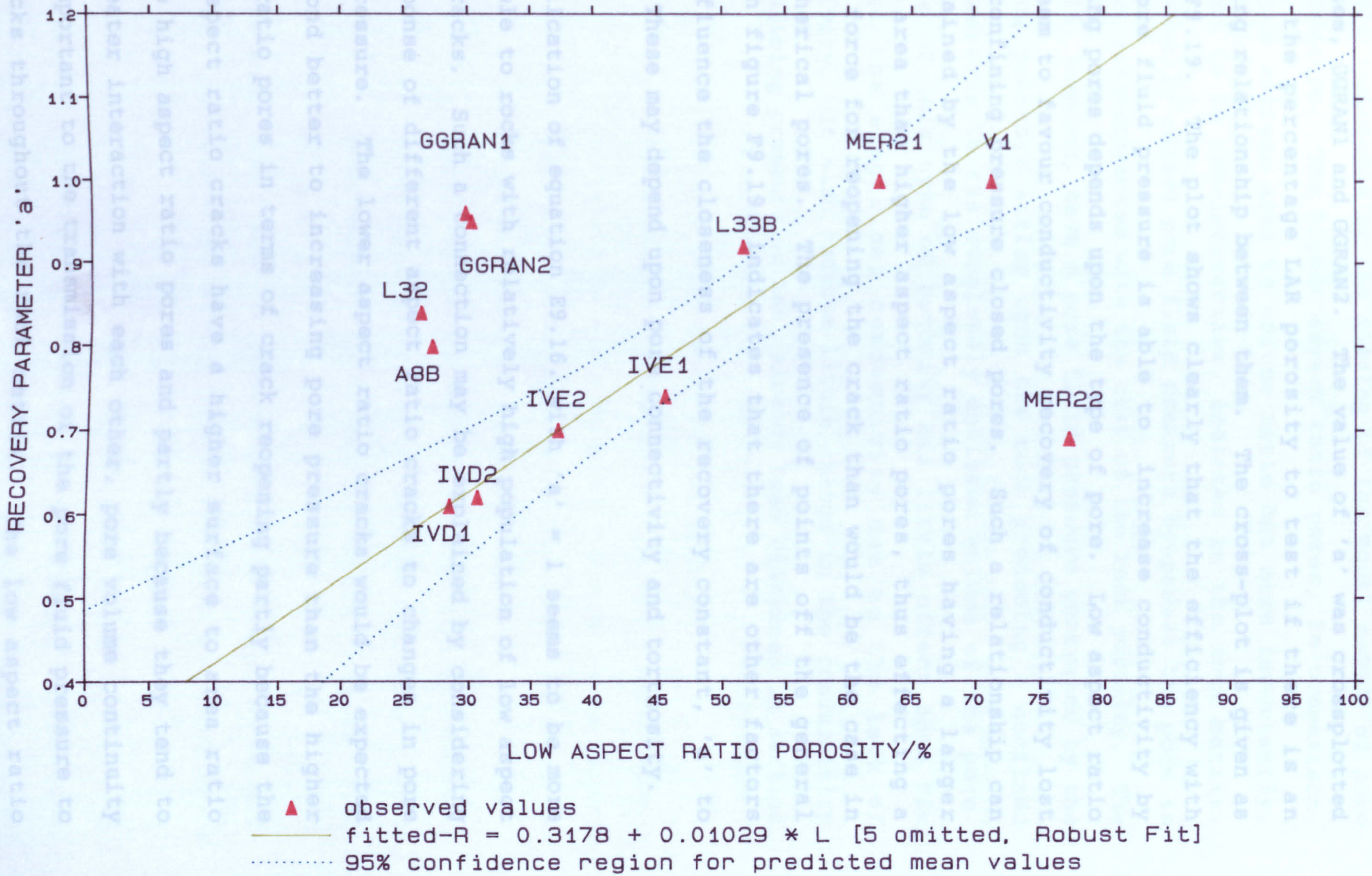
where the constant, a, is less than or equal to one. If a = 1 then the effect of pore fluid pressure is equal and opposite to the confining pressure. In reality the value of the constant, a, varies from rock to rock and for different measurable quantities such as electrical conductivity or acoustic P-wave velocity. The values of 'a' for the electrical conductivity variation measured for each of the rocks studied in this work have been calculated by comparison of the effect of pore fluid pressure and confining pressure upon the magnitude of the electrical conductivity changes. The values of this constant are given in table T9.5.

TABLE T9.5. VALUES OF THE CONSTANT 'a' FOR ALL THE ROCK SAMPLES STUDIED

AMPHIBOLITE	
IVD1	0.61
IVD2	0.62
GRANULITE	
IVE1	0.74
IVE2	0.70
ORTHOPYROXENITE	
MER21	1.00
MER22	0.69
AMPHIBOLITE GNEISS	
A8B	0.80
GRANODIORITE	
L33B	0.92
AMPHIBOLITE GNEISS	
L32	0.84
GRANITE	
V1	1.00
GRAPHITE GRANULITE	
GGRAN1	0.96
GGRAN2	0.95

The samples in table T9.5. are paired in terms of similar, or the same mineralogy. In all cases but one there is close agreement between the individual samples in each pair.

FIGURE F9.19. CROSSPLOT OF THE PORE FLUID PRESSURE SCALING RECOVERY PARAMETER 'a' AGAINST THE LOW ASPECT RATIO POROSITY PERCENTAGE



The samples with 'a' approaching unity are the two acidic rocks, the orthopyroxenite, MER21, and the pair of graphite garnet granulites, GGRAN1 and GGRAN2. The value of 'a' was crossplotted against the percentage LAR porosity to test if there is an underlying relationship between them. The cross-plot is given as figure F9.19. The plot shows clearly that the efficiency with which pore fluid pressure is able to increase conductivity by re-opening pores depends upon the type of pore. Low aspect ratio pores seem to favour conductivity recovery of conductivity lost due to confining pressure closed pores. Such a relationship can be explained by the low aspect ratio pores having a larger surface area than higher aspect ratio pores, thus effecting a greater force for reopening the crack than would be the case in more spherical pores. The presence of points off the general trend in figure F9.19. indicates that there are other factors that influence the closeness of the recovery constant, 'a' to unity. These may depend upon pore connectivity and tortuosity.

The application of equation E9.16. with 'a' = 1 seems to be more applicable to rocks with relatively high population of low aspect ratio cracks. Such a connection may be explained by considering the response of different aspect ratio cracks to changes in pore fluid pressure. The lower aspect ratio cracks would be expected to respond better to increasing pore pressure than the higher aspect ratio pores in terms of crack reopening partly because the lower aspect ratio cracks have a higher surface to area ratio than the high aspect ratio pores and partly because they tend to have greater interaction with each other, pore volume continuity being important to the transmission of the pore fluid pressure to the cracks throughout the rock matrix. The low aspect ratio

cracks, when closed by a confining pressure, may be treatable as completely reopenable by the raising of the equal amount of pore fluid pressure. The higher aspect ratio pores, in comparison, have a low surface area to volume ratio but more importantly, they tend to be non-interactive, isolated in the rock matrix. The externally raised pore fluid pressure by-passes the pore as it is not interconnected with the rest of the rock porosity. The pore will, however, have a pore fluid pressure generated by the confining pressure acting upon the rock producing a confined fluid pressure in the completely enclosed volume of the pore. The size of this type of porosity has little effect upon the value of the electrical conductivity due to the lack of connectivity. It also induces little change in the conductivity with confining pressure as has already been discussed in section 9.1.

9.3. THE EFFECT OF TEMPERATURE UPON THE ELECTRICAL CONDUCTIVITY OF SATURATED ROCK SAMPLES HELD AT HIGH CONFINING AND PORE PRESSURES

9.3.1. INTRODUCTION

It has already been mentioned that the major aim of this work was the extension of electrical conductivity measurements upon saturated rocks to temperatures higher than 250°C. This particular goal has been achieved inasmuch as measurements have been possible up to 900°C at raised confining and pore pressures. The results are given as figures F8.27. to F8.62. in three different ways:

1. The initial, pre-temperature transient drop, electrical conductivity variation as a function of temperature in degrees C at a pore pressure of 0.1 GPa (Figures F8.27. to F8.38.).
2. The subsequent repeatable electrical conductivity variation as a function of temperature in degrees C at pore pressures of 0.08, 0.10 and 0.18 GPa (Figures F8.39. to F8.50).
3. The repeatable electrical conductivity variation at various pore fluid pressures plotted against 1000/Temperature in degrees kelvin (Figures F8.51. to F8.62.).

The data have been used to generate various further parameters including the activation energy, E_o, which is given in figures F9.21. to F9.32.

Discussion of the effect of temperature upon the electrical conductivity of saturated rocks at high pressures will be given

in the order of the presentation of the data.

9.3.2. INITIAL TEMPERATURE VARIATIONS

The initial temperature variation is that electrical conductivity variation that takes place with the initial raising of temperature. There is an increase in the electrical conductivity of saturated rocks with increase of temperature. This is mainly associated with the increase in the electrical conductivity of the saturating pore fluid as temperature rises. The samples were all left at high temperatures for up to 15 hours immediately after the initial temperature rise. The electrical conductivity of the rock dropped during this period. The drop has been termed the temperature transient drop because it takes place at raised temperatures held for a period of time and because it only occurs once for a set of confining pressure, pore pressure and temperature values. The drop has been observed by a few workers (Lee, 1977; Evans, 1980; Jackson, 1984) and is thought to be well understood. The effect has been thoroughly discussed in chapter 2 and section 9.1. Briefly, it may be seen as an exacerbation of the microstructural changes caused by the stressing of a rock with applied confining and pore pressures. Raising the temperature disturbs the balance between the rock matrix strength and the applied pressures that was achieved at the end of the experimental changes in pressure. The microstructure, and hence electrical conductivity, changes until a new balance is achieved. The temperature transient drop has been labelled on all the initial temperature variation curves (figures F8.27. to F8.38.).

Figures F8.27. to F8.30. show the initial electrical conductivity

variation with temperature for confining pressures of 0.2 GPa and pore fluid pressure of 0.18 GPa. The electrical conductivity increases swiftly with increase in temperature. It was recognised before the experiments began that the repeatable electrical conductivity variation was much more important data than the pre-transient data. The time spent in raising the temperature initially was minimised in order to maximise the time spent on making post-transient measuurements. One of the side-effects of this policy was that the internal core temperature of the sample lagged behind the measured temperature at the thermocouple tip on the low temperature initial pre-transient data. This effect is responsible for the low temperature concavity seen in most of the initial temperature variation data. The data is to be judged as descriptive of a general trend rather than being strictly accurate. The accuracy has recovered by 500^{o}C. The concavity was reduced in the later runs by controlling the rate of temperature rise with a three term temperature controller. The concavity in the later runs is reduced very much but is still present.

Figures F8.31. and F8.33. to F8.38. show the initial electrical conductivity variation for confining pressure of 0.2 GPa and a pore pressure of 0.1 GPa. They also show the initial electrical conductivity variation when the pore pressure had been raised to 0.18 GPa after all the temperature measurements at pore pressures of 0.1 GPa had been made. There is also a concavity in this data due to the same experimental method as has already been discussed in the last paragraph. The shape of the initial variation and the transient drop occur for the same reasons that have already been discussed. Once all the repeatable post-transient

electrical conductivity measurements had been done at a pore pressure of 0.1 GPa the system was returned to laboratory temperatures. The pore pressure was increased to 0.18 GPa and the initial electrical conductivity variation with temperature at this new value of pore fluid pressure was measured. A temperature transient change (increase) would be expected as one of the set of P_c, P_p and T values have been changed. The increase in the pore pressure of a rock tends to increase the electrical conductivity of the rock by opening pores. This process is not immediate. Increasing the temperature enables this process to progress a little further until the rock is again in a stable state. Leaving the sample at raised temperatures, having increased the pore pressure produced a small temperature transient increase as expected. The curve is included in these figures as an initial variation because it is the initial variation of the electrical conductivity for a new set of P_c, P_p, T values and behaves as such, laying aside the point that the curve was obtained after repeatable measurements were made on the rock at pore pressures of 0.1 GPa and that these variations are very similar to the repeatable post-transient measurements for that reason.

The initial electrical conductivity variations are of little use other than in illustrating the general behaviour and the temperature transient drop. It is the repeatable measurements that hold most of the information concerning the behaviour of saturated rocks that are under long term pore pressure, confining pressure and temperature conditions in the Earth's crust.

9.3.3. POST-TEMPERATURE TRANSIENT VARIATIONS OF CONDUCTIVITY WITH TEMPERATURE

Figures F8.39. to F8.50. show the repeatable temperature variations for all 12 samples. All samples were measured at a confining pressure of 0.2 GPa. The range of values of pore pressures are shown on the individual figures. All the samples gave good repeatable post-temperature transient variations of electrical conductivity.

The repeatable temperature variation of electrical conductivity falls into two different types:

1. Electrical conductivity increases with temperature up to temperatures up to 900 ^{o}C.
2. Electrical conductivity initially increases with temperature but subsequently reaches a maximum value after which the electrical conductivity begins to rapidly decrease.

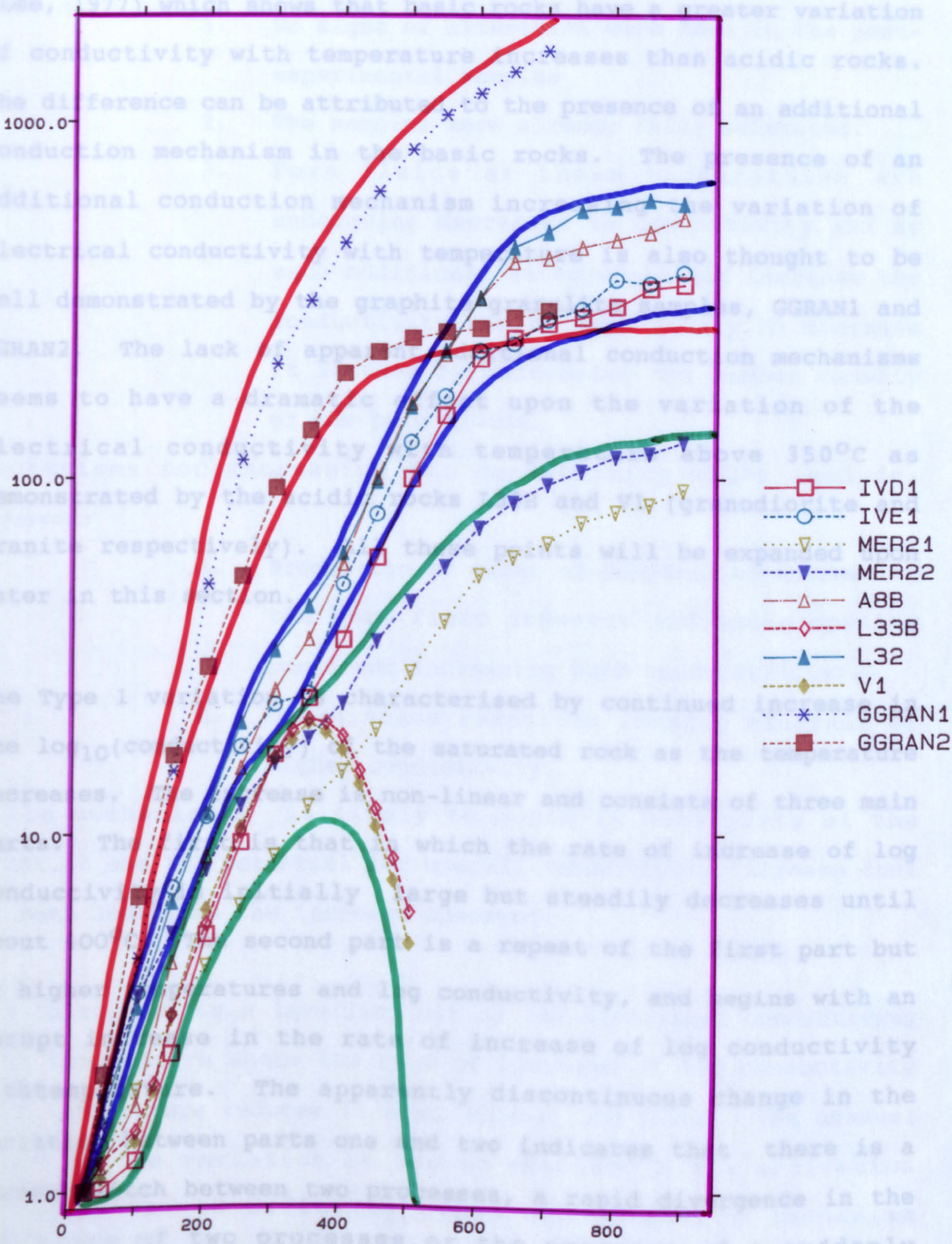
The Type 1 variation is followed by all the more basic rocks studied: Basic granulites; IVD1,IVD2,
Basic amphibolites; IVE1,IVE2,
Orthopyroxenites; MER21,MER22,
Amphibolite gneisses; A8B,L32,
Garnet graphite granulites; GGRAN1,GGRAN2.
These are shown in figures F8.39. to F8.45., F8.47., F8.49., F8.50.

The Type 2 variation is followed by both of the acidic rocks studied: Granodiorite; L33B,
Granite; V1.
These are shown in figures F8.46. and F8.48.

There is a large increase in the electrical conductivity with temperature. Variations of several orders of magnitude occur. The electrical data has been expressed in terms of log to the base 10 to enable better presentation of the data.

All the repeatable temperature variations have been collected and are presented as data normalised to the value of their conductivity at 25^{o}C in figure F9.20. This type of graph is very useful in comparing the rate of change of each individual variation. The temperature variations of electrical conductivity with temperature can be split into three sections which are coloured separately in figure F9.20. The two samples showing the greatest variation of electrical conductivity with temperature are the graphite granulites, GGRAN1 and GGRAN2 (figures F8.49 and F8.50, and shown in the red section of figure F9.20). These variations will be discussed at greater length in section 9.3.8. The basic amphibolites, IVD1 and IVD2; basic granulites, IVE1 and IVE2; and amphibolite gneisses, A8B and L32 make up another section (shown in blue in figure F9.20) with large electrical conductivity variations at all values of temperature and increasing to high electrical conductivities at high temperatures. The third section (coloured in green in figure F9.20) includes those rocks showing an initial lower variation of electrical conductivity with temperature compared with those rocks in the other two sections. This last section includes the acidic rocks such as the granodiorite, L33B and the granite, V1. These samples show surprising reductions in conductivity with temperature at temperatures greater than about 350^{o}C. The section also includes the orthopyroxenites, MER21 and MER22 which

FIGURE F9.20. THE VARIATION OF CONDUCTIVITY NORMALISED TO THAT AT 25 DEGREES C WITH TEMPERATURE FOR DIFFERENT ACIDIC AND METABASIC ROCKS



show a basic rock type variation but with a much lower gradient than the basic rocks in the other two sections. The results shown in figure F9.20 agree well with previous work up to 250°C (Lee, 1977) which shows that basic rocks have a greater variation of conductivity with temperature increases than acidic rocks. The difference can be attributed to the presence of an additional conduction mechanism in the basic rocks. The presence of an additional conduction mechanism increasing the variation of electrical conductivity with temperature is also thought to be well demonstrated by the graphite granulite samples, GGRAN1 and GGRAN2. The lack of apparent additional conduction mechanisms seems to have a dramatic effect upon the variation of the electrical conductivity with temperature above 350°C as demonstrated by the acidic rocks L33B and V1 (granodiorite and granite respectively). All these points will be expanded upon later in this section.

The Type 1 variation is characterised by continued increase in the $\log_{10}$(conductivity) of the saturated rock as the temperature increases. The increase is non-linear and consists of three main parts. The first is that in which the rate of increase of log conductivity is initially large but steadily decreases until about 400°C. The second part is a repeat of the first part but at higher temperatures and log conductivity, and begins with an abrupt increase in the rate of increase of log conductivity withtemperature. The apparently discontinuous change in the variation between parts one and two indicates that there is a sudden switch between two processes, a rapid divergence in the dominance of two processes or the presence of a suddenly switching bistable process. It has also been suggested that the

discontinuity marks the onset of dehydration reactions releasing water (Hinze, personal communication). This mechanism is unlikely as:

1. No signs of alteration were seen in the post-experimental samples.
2. The samples were already fully saturated.
3. Pore fluids at these temperatures are undergoing decreases in conductivity and as such additional water would not increase the conductivity but be more likely to decrease it further by decreasing the number density of the pore fluids.

Mechanisms for increasing the conductivity might include, however:

A. Production of water of dehydration increasing the pore fluid pressure and hence opening pores and increasing bulk conductivities.
B. Dehydration reactions forming minerals of higher conductivity.

While mechanism A is likely to occur in some parts of the crust,it has the potential for overall conductivity increase that is much less than the increase observed.

The third part is a levelling off of the electrical conductivity with temperature where the rate of increase of log conductivity with temperature reduces to a relatively low level. The unusual shape of the variation is due to changes in the activation energies due to differing types of conduction mechanism dominating the bulk rock conductivity at different temperatures.

This effect will be discussed at greater length later in this section.

The increase in the electrical conductivity with temperature noted by **Lee (1977)** and **Jackson (1984)** in their experiments up to 250°C continues up to around 400°C. Up to this temperature the rate of change of log conductivity with temperature gradually decreased. At around 400°C the rate of change of log conductivity with temperature rapidly increased and then gradually decreased as temperature was further raised until temperatures of around 700°C were reached. The behaviour of the log electrical conductivity at the higher temperatures was very variable but generally included a flattening out of the variation.

The Type 2 variation is characterised by an increase in the value of log conductivity with temperature followed by a decrease in log conductivity with temperature. The variation is non-linear and can be described in two main parts. The first part consists of an increase in log conductivity with temperature that progressively lessens until a critical temperature is reached at which there is instantaneously no increase of log conductivity with temperature. The critical temperature occurs at around 350°C and is independent of rock type in the two rocks of this type that were studied. The second part consists of a decrease of log conductivity with temperature. The rate of decrease of log conductivity becoming greater as the temperature increases. The post-critical temperature variation is greater for those rocks held at a lower pore fluid pressure. The continuity of the variation through the critical temperature indicates that the

variation is due to either the continuous variation of some process controlling the electrical conductivity of the rocks or the gradual switching between more than one process.

The increase in conductivity over the full temperature range is very large for both Type 1 and Type 2 variations. It ranged from two to just over three orders of magnitude. The observed conductivities can be applied to various generalised depths within the crust by estimating the temperatures present at these levels. These figures are given for both Type 1 and Type 2 variations in table T9.6.

TABLE T9.6. CONDUCTIVITIES AT NOMINAL CRUSTAL LEVELS FROM MEASURED ELECTRICAL CONDUCTIVITY/TEMPERATURE VALUES

LEVEL	CONDUCTIVITY TYPE 1 $S\ m^{-1}X10^{-3}$	CONDUCTIVITY TYPE 2 $S\ m^{-1}X10^{-3}$	CONDUCTIVITY MAGNETO-TELLURIC OBSERVED* $S\ m^{-1}X10^{-3}$
UPPER CRUST	0.02 - 10	0.03 - 5	0.05 - 0.1
UPPER LOWER CRUST	2 - 50	1 - 5	10 - 50
LOWER LOWER CRUST	30 - 150	0.04 - 5	1 - 10

* FROM JONES, 1981; BEAMISH, 1986a,b

The above table was generated by taking the average values for the rocks exhibiting each type of behaviour. The table assumes the rocks in the crust to be fully saturated with 0.5 M NaCl solution just as those samples used in the laboratory experiments. There is likely to be variation away from these categories due to:

1. A different concentration or type of electrolyte being present in the pores of deep crustal rocks.
2. The pores being partially saturated or liquid-free.
3. Variations in the local heat flow causing local areas where the temperature of the crust at all levels is raised or lowered from the usual range.

Basic rocks saturated with 0.5 M NaCl solutions such as those undergoing type 1 variations in this work, have electrical conductivities consistent with those obtained from magneto-telluric measurements of the electrical conductivity at various depths within the crust. This applies in particular to the upper crust and the high conductive zone of the upper lower crust.

The upper crust field values are explainable in terms of either type 1 or type 2 saturated rocks. The rocks of the upper crust can be expected to be saturated with a brine, the concentration of which is approximately that of seawater. Both type 1 and type 2 saturated rocks have a bulk conductivity dominated by electrolytic conduction but the basic type 1 rocks have a range of electrical conductivities shifted to higher conductivities by an additional conduction mechanism.

The upper part of the lower crust have measured field values explainable only in terms of type 1 basic saturated rocks. Table T9.6. clearly shows that the type 1 laboratory measured values cover the whole range of observed field values, whereas the conductivity range of laboratory measured type 2 acidic saturated rocks falls entirely below the conductivity range of observed field values. Clearly some additional mechanism must be raising

the the bulk saturated rock conductivity of the type 1 basic rocks and doing so to a much greater extent than was the case at the upper crustal temperatures.

The lower part of the lower crust has a range of magneto-telluric conductivity measurements significantly lower than those measured in the laboratory on type 1 basic rocks but in agreement with the range of conductivities obtained from laboratory measuremnts upon saturated type 2 acidic rocks. It is very unlikely that the lower part of the lower crust consists of such acidic rocks from arguements based upon lower lower crustal stability, density inversions, partial melting and mechanisms for crustal formation. The apparent agreement of laboratory based saturated acidic rock conductivities with the observed field data compared with the higher values from laboratory studies on saturated type 1 rocks is more likely to be caused by either:

1. The magnetotelluric field data being underestimated due to an inherent weakness in MT inversion techniques that confuse thin highly conducting layers with thicker, moderately high conductivity layers.

and/or;

2. The laboratory conductivity nmeasurements being unrealistically high due to the presence of a greater amount of water than may be the case in an anhydrous lower lower crust.

9.3.4. INVERSE TEMPERATURE VARIATIONS

The temperature variations can be better understood if presented in a slightly different way. The variation of electrical conductivity with temperature can be described by an Arrhenius equation providing that the charge carrier density and mobility remains constant with temperature increase. Such a variation gives a $\log_e$(conductivity)/1000/T(kelvin) graph which is linear. The gradient of this graph is equal to $-E_o/k$ where E_o is the activation energy and k is the Boltzmann constant. The activation energy is the parameter which defines the rate of change of the electrical conductivity of a medium with temperature providing the mechanism, amount and mobility of the charge carriers remains constant with temperature. I have called this method of plotting the 'inverse' temperature variation. A full and more mathematical description of the method has been discussed in full in chapter 2.

The Type 1 variations have been replotted against 1000/T(kelvin). These variations are shown in figures F8.51. to F8.57., F8.59. and F8.61. to F8.62. The Type 2 variations have also been replotted and are given as figure F8.58. and F8.60. the figures show clearly that there are multiple linear portions. Multiple activation energies can be inferred which in turn require the conduction mechanism to have changed in some way. In some cases this is due to the increase in temperature altering the charge carrier density or mobility of a previously existing mechanism such as is the case with the Type 2 variations: the temperature increase produces an isobaric reduction of the charge carrier density by thermal expansion and association effects. In other cases the change in activation energy may be due to an

another type of conduction mechanism adding to or replacing a previously important mechanism.

Most of the Type 1 variation rock samples have an 'inverse' variation consisting of three linear phases corresponding to the three parts existing on the temperature variation as previously descibed. There is a low temperature phase with a certain negative gradient, a medium temperature phase with a larger negative gradient and a third, high temperature, phase with a small negative gradient. In some rocks the first phase may be preceded by a phase characterised by a very low gradient and that does not exist above 100°C.

The Type 2 'inverse' temperature variations show a phase of very low negative gradient up to around 100°C corresponding to that pre-phase found in some of the Type 1 rocks. A second phase of higher negative gradient ends rapidly becoming less steep, reaching a maximum value and swiftly gaining a large positive gradient.

The low initial gradient of the type 2 rocks inverse temperature variations compared with that of the type 1 rocks corresponds with lower rates of increase of conductivity in the type 2 rocks compared with those of the type 1 rocks. This result is in agreement with Lee (1977) and Jackson (1984) who found that the electrical conductivity of acidic rocks generally increase at a slower rate than the more basic rocks.

9.3.5. ACTIVATION ENERGY/TEMPERATURE GRAPHS

Reference to chapter 2 shows that the gradient of a $\log_e$(conductivity)/1/T(kelvin) graph gives a value equal to $-E_o/k$ from which a value of the activation energy, E_o, can be calculated. The $\log_{10}$(conductivity)/1000/T(kelvin) graphs shown in figures F8.51. to F8.62. have been processed using RS/1 software to give $E_o/T(^oC)$ for each of the variations. These graphs are shown as figures F9.21. to F9.32. A rolling algorithm was used to obtain the gradient from two consecutive points over the range of 1000/T(kelvin) present in each of the source graphs. Calculation of gradient from two consecutive points amplifies small errors in the original data and produces a greater amount of scatter in the E_o data than would otherwise be expected. The scatter can be improved by using more than two consecutive points to calculate the gradient. The use of such a calculation 'window' inevitably degrades the temperature resolution of the final E_o/T graph by causing values of activation energy that lie close to a change in activation energy to be affected by the other gradient. In effect the calculation for the gradient 'sees' a part of another gradient and is therefore slightly biased towrds the other gradient. The effect can be seen in the E_o/T graphs as a rounding of the response near activation energy changes. The effect is minimal if the number of datapoints on each linear portion in the source data is greater than at least two times the number used in the calculation of the gradient at a specific point. Thus if the gradient were calculated over three points and the linear portion had seven points lying in a well defined straight line there would be five gradients in good mutual agreement. In this work the number of data-points on some small linear portions of curve are such that the portion is

recognisable by eye but low enough for the activation energy variation calculated by the computer to show rounded shoulders associated with this effect. An example of this can be seen in figures F9.21. and F8.51. The linear portion recognised by eye in figure F8.51. between 1000/T(kelvin) values of 1.1 and 1.4 translates into figure F9.21. as the four points occurring around E_o ~ 0.45 - 0.50 eV. The two side points are artificially lowered by the method of calculation. The only way to improve the variation is to increase the density of the source datapoints.

The calculating algorithm proceeded by:

1. calculating the gradient of the source graph at consecutive points by stepping through 1000/T(kelvin) values.
2. converting the value of $-E_o/k$ so obtained to give an initial value of E_o.
3. correcting for the $\log_{10}$(conductivity) source data to give the value which $\log_e$(conductivity) would have given.
4. correcting for the 1000/T(kelvin) source data to give the value which 1/T(kelvin) would have given.
5. scaling the value of E_o to be in electron-volts.
6. plotting the obtained value of activation energy against temperature in degrees centigrade. The temperature was calculated by taking the average of the 1000/T(kelvin) values used in the gradient calculation and converting into temperature in degrees centigrade.

FIG. F9.21. IVD1TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED BASIC AMPHIBOLITE

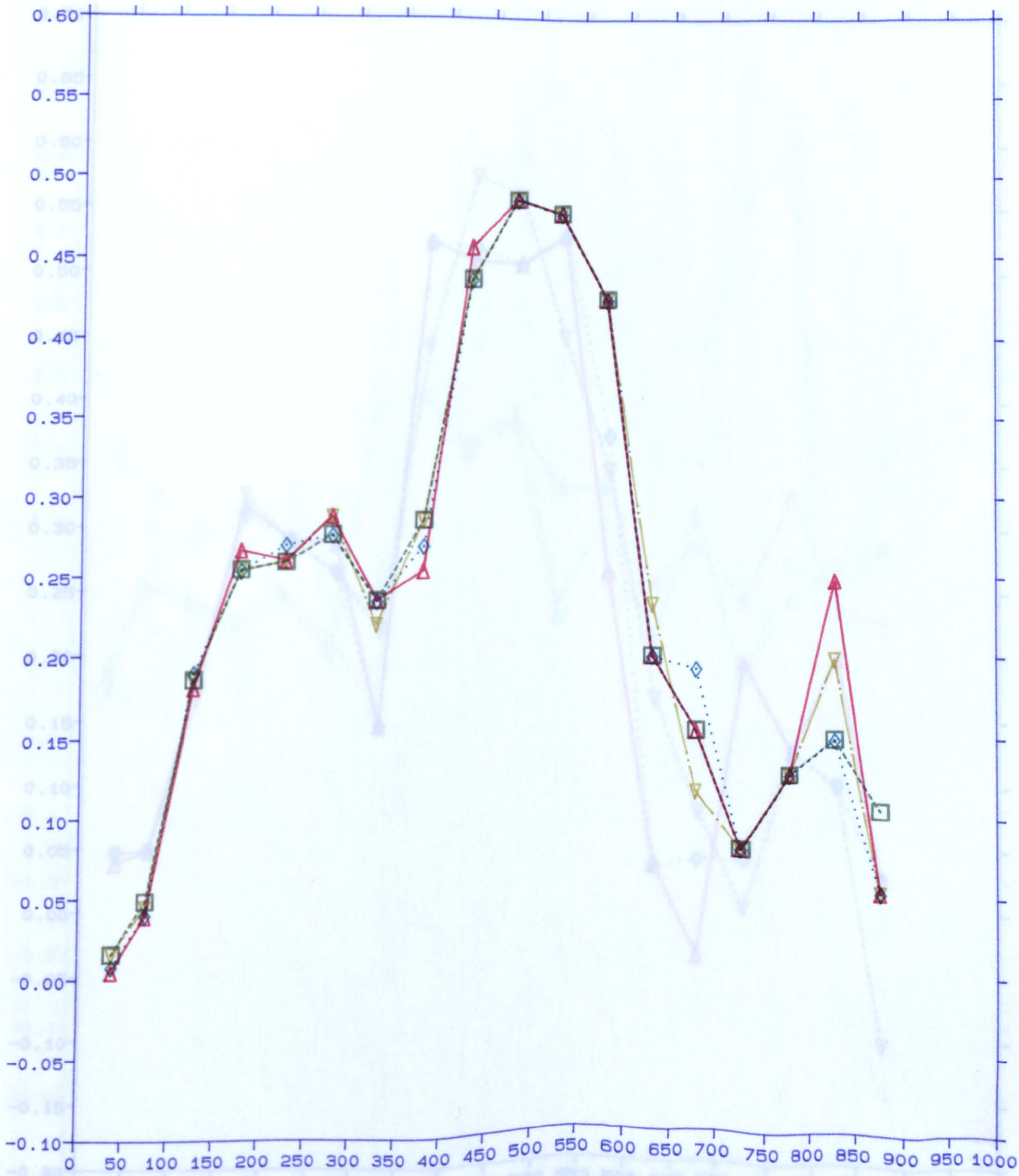


FIG. F9.22. IVD2TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR DISTILLED WATER SATURATED BASIC AMPHIBOLITE

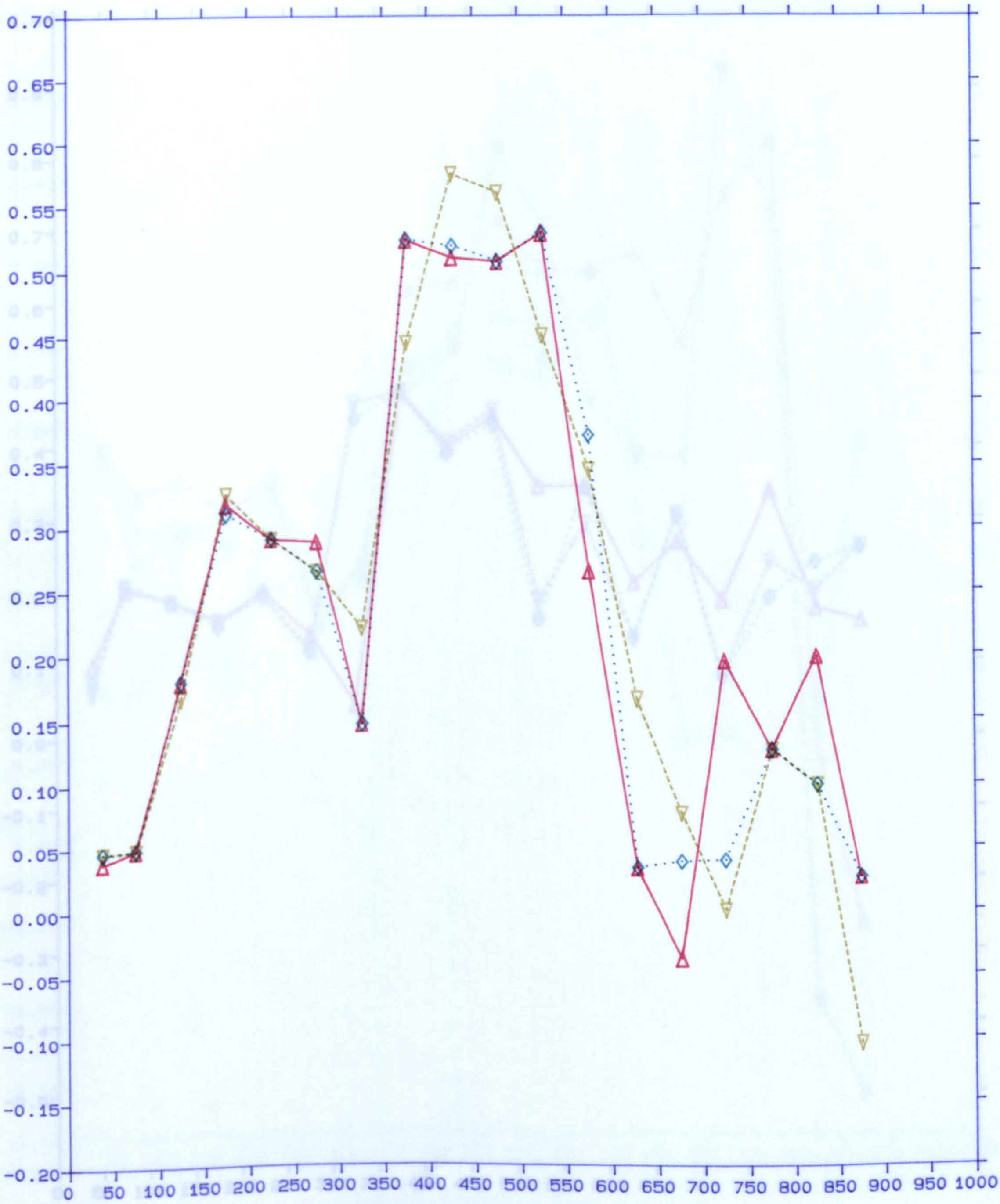


FIG. F9.23. IVE1TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED BASIC GRANULITE

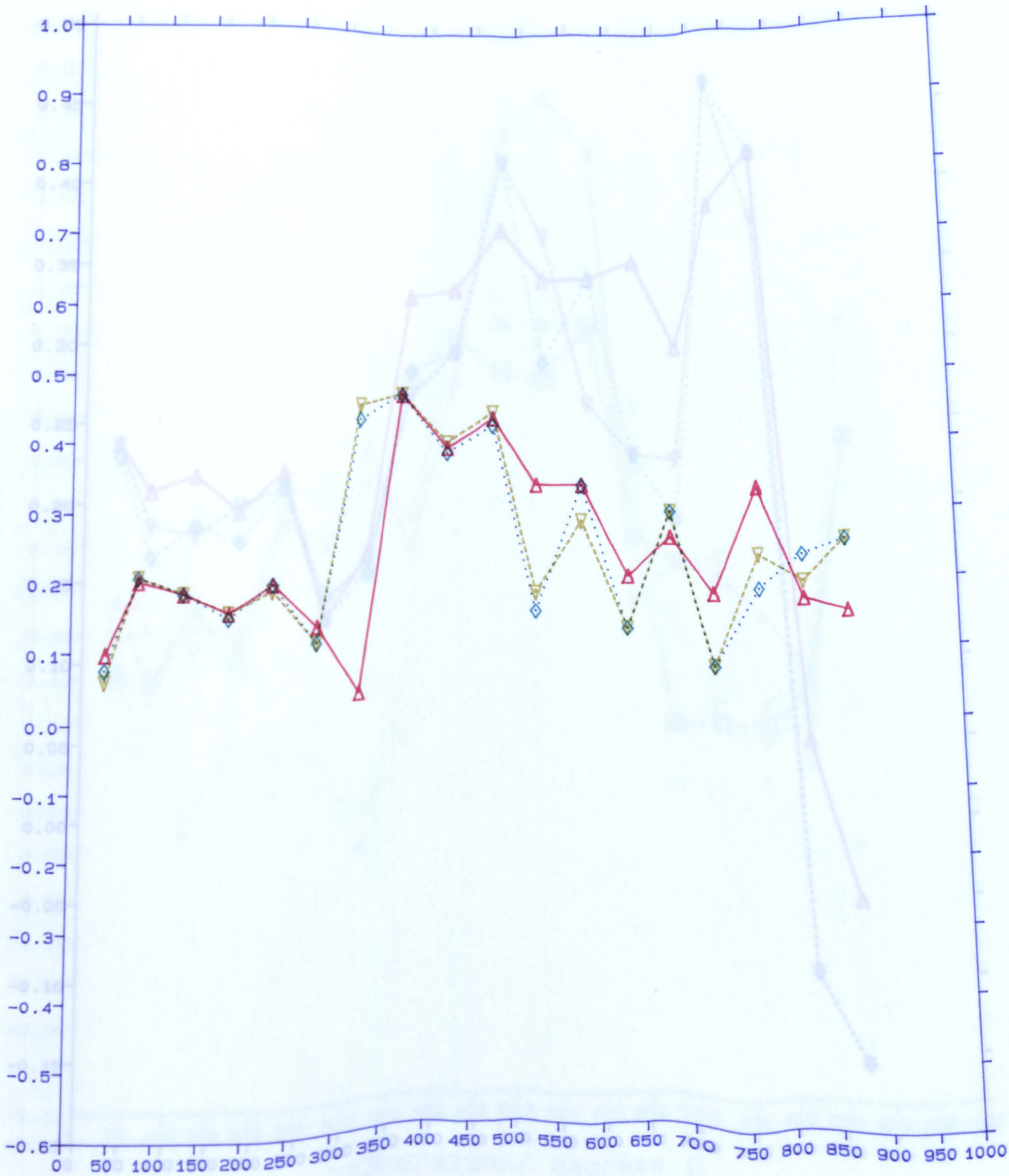


FIG. F9.24. IVE2TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR DISTILLED WATER SATURATED BASIC GRANULITE

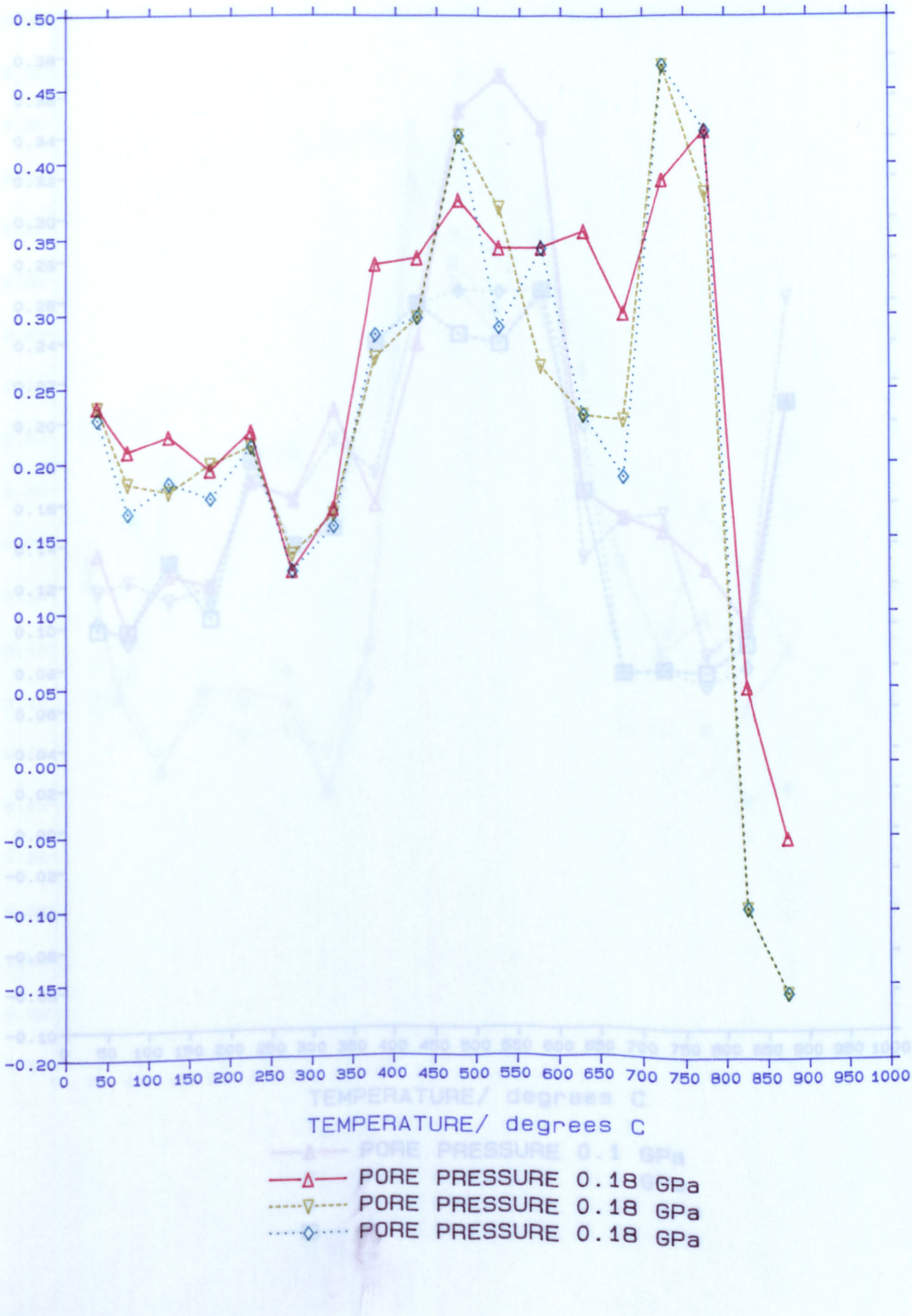


FIG. F9.25. MER21TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED ORTHOPYROXENITE

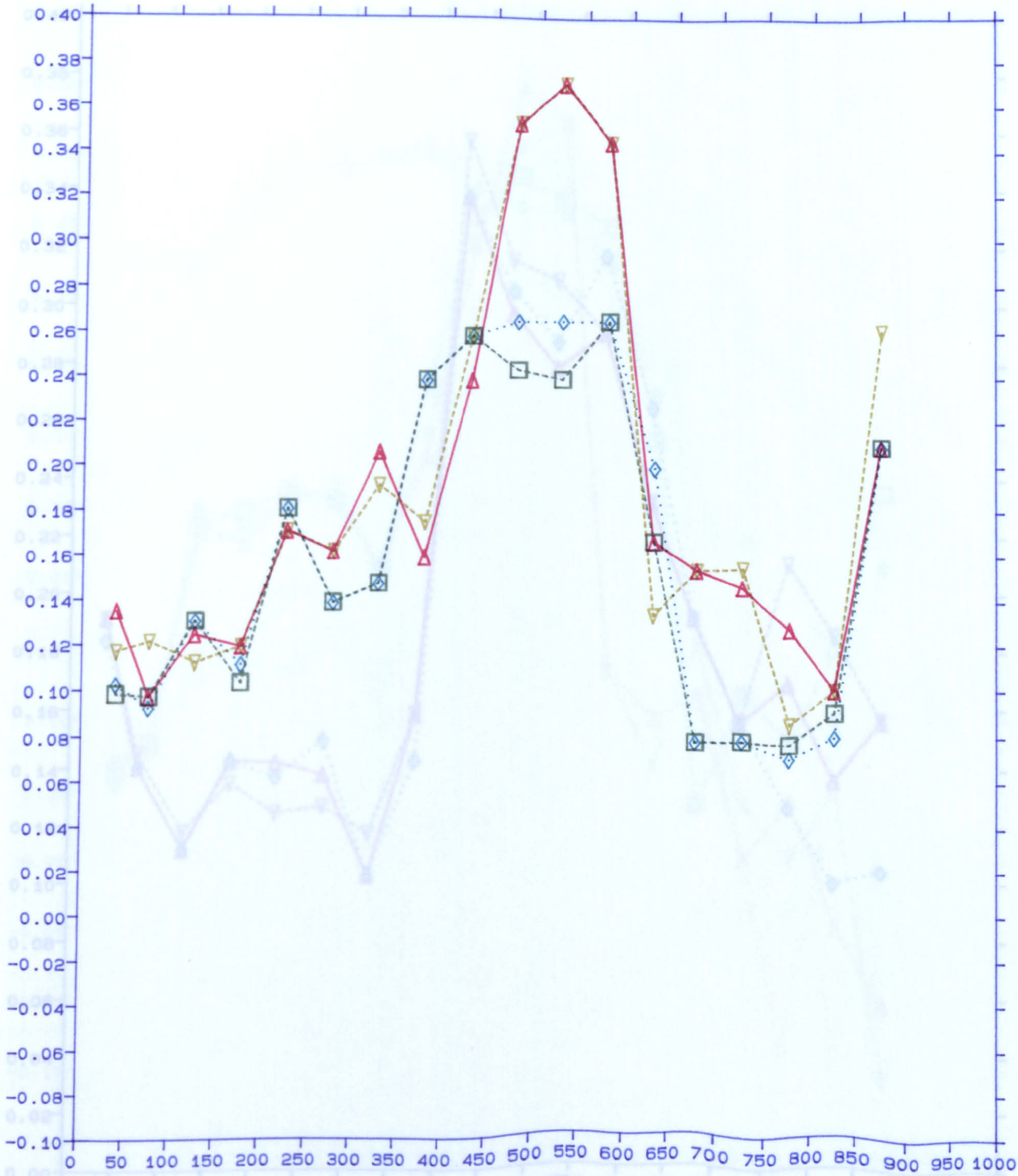


FIG. F9.26. MER22TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR DISTILLED WATER SATURATED ORTHOPYROXENITE

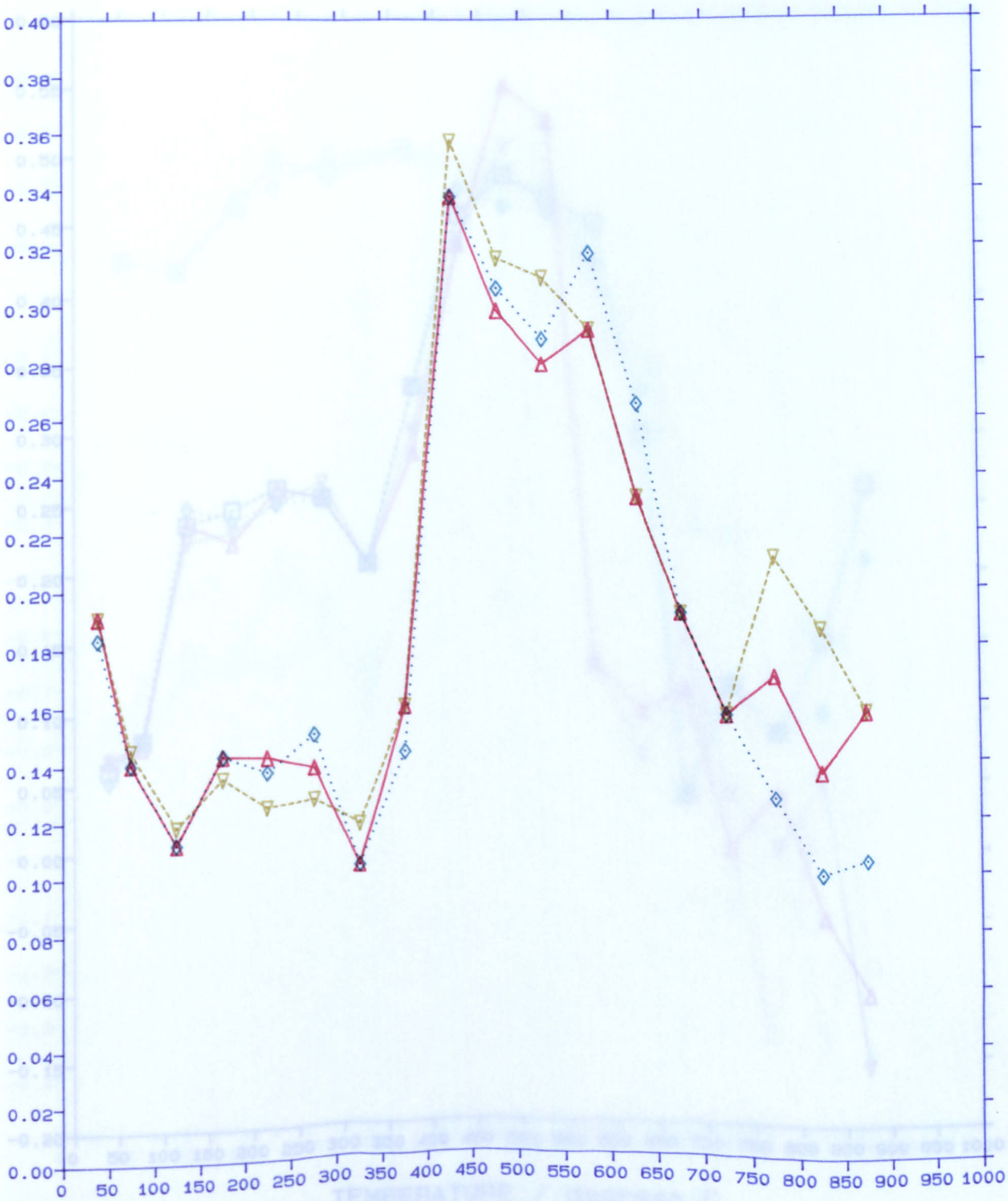


FIG. F9.27. A8BTA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

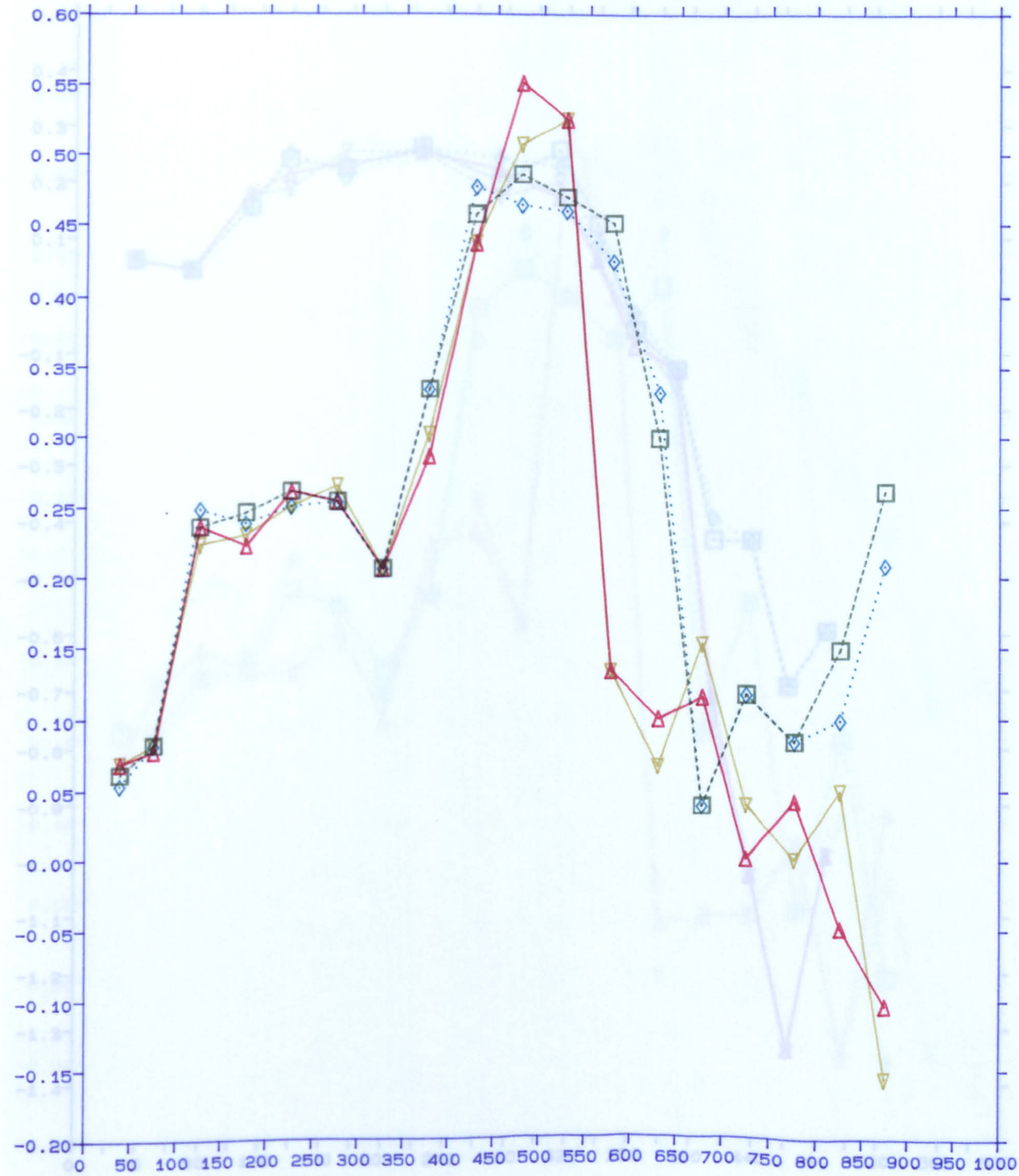


FIG. F9.28. L33BTA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED GRANODIORITE

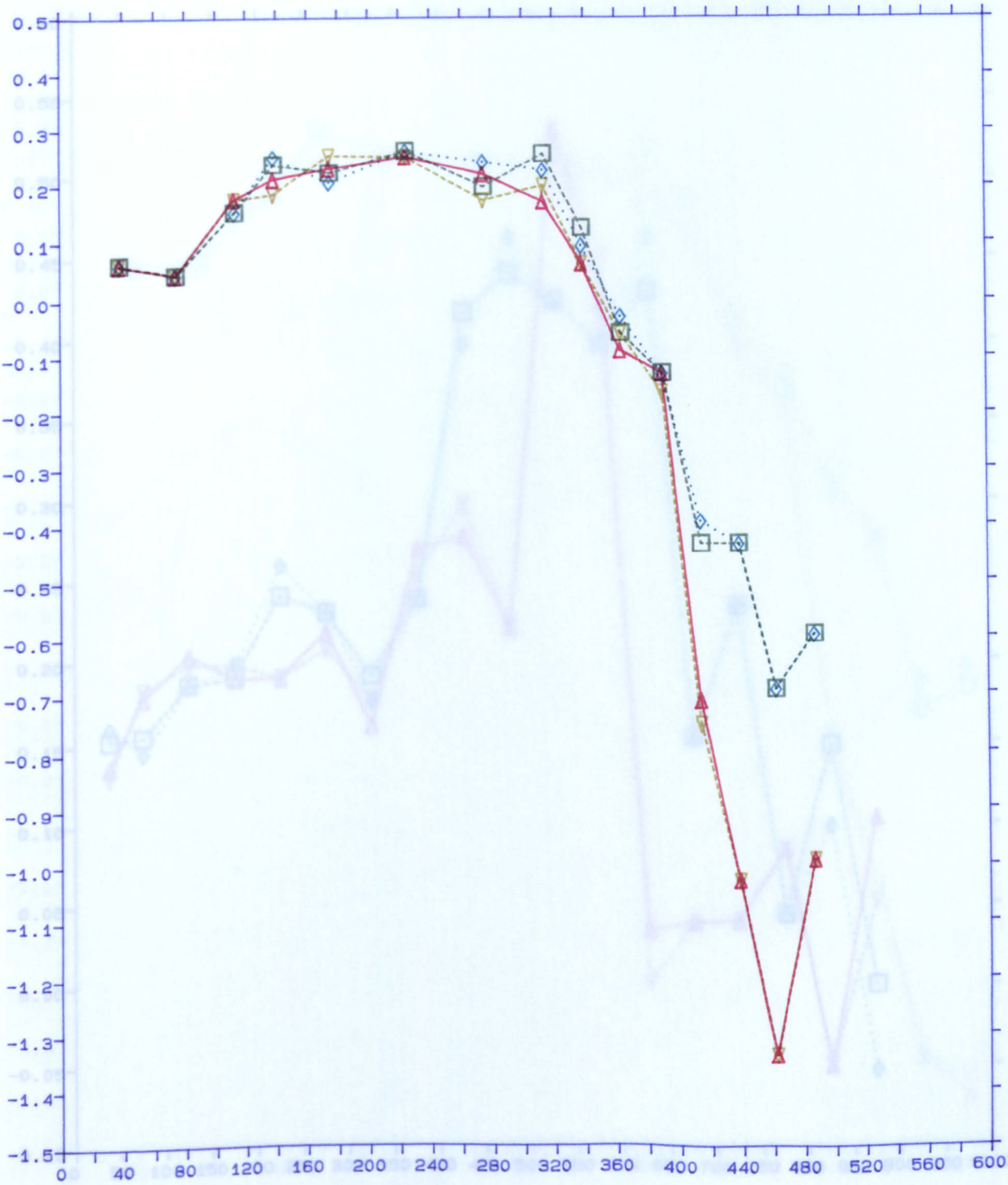


FIG. F9.29. L32TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED AMPHIBOLITE GNEISS

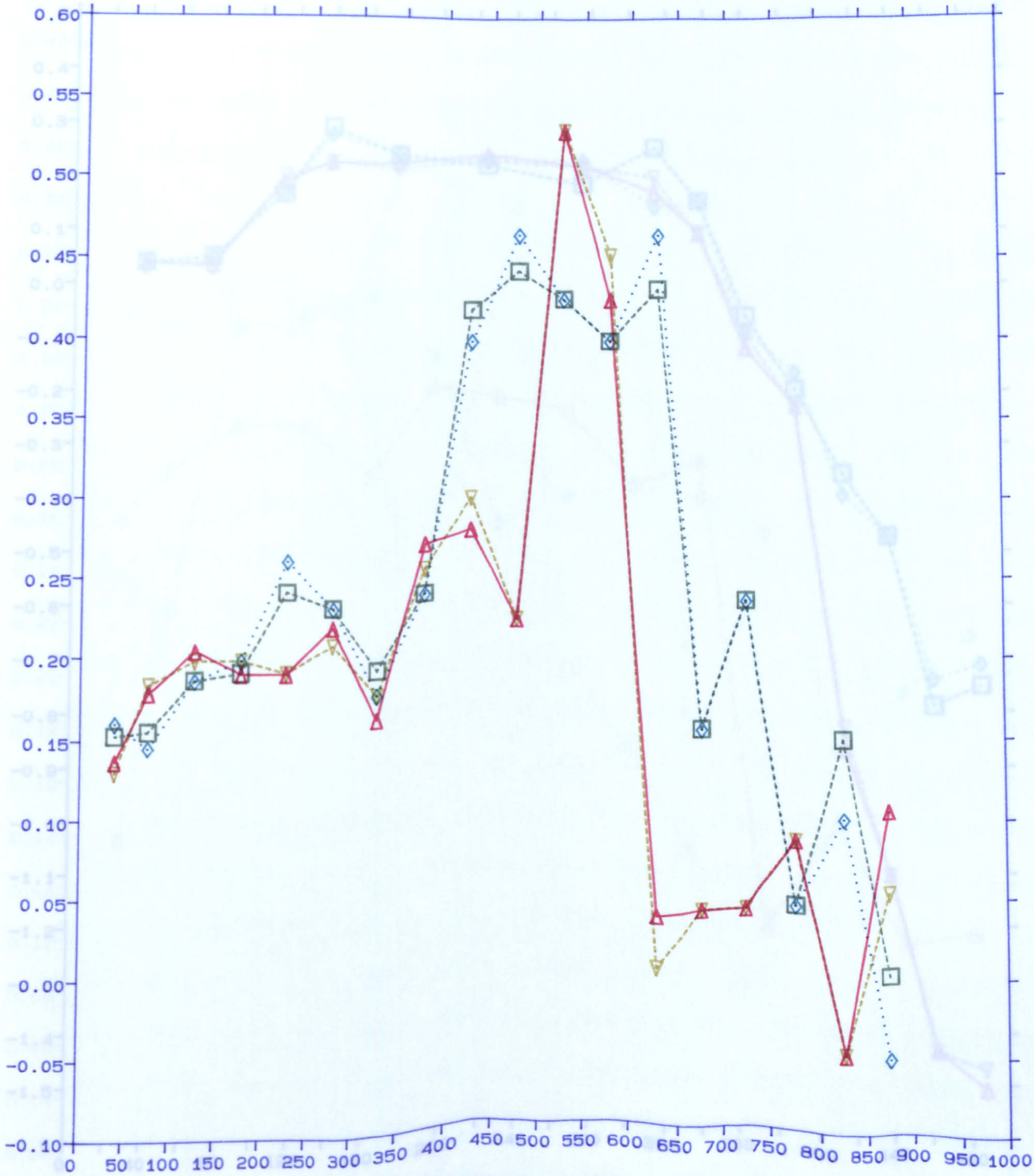


FIG. F9.30. V1TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED GRANITE

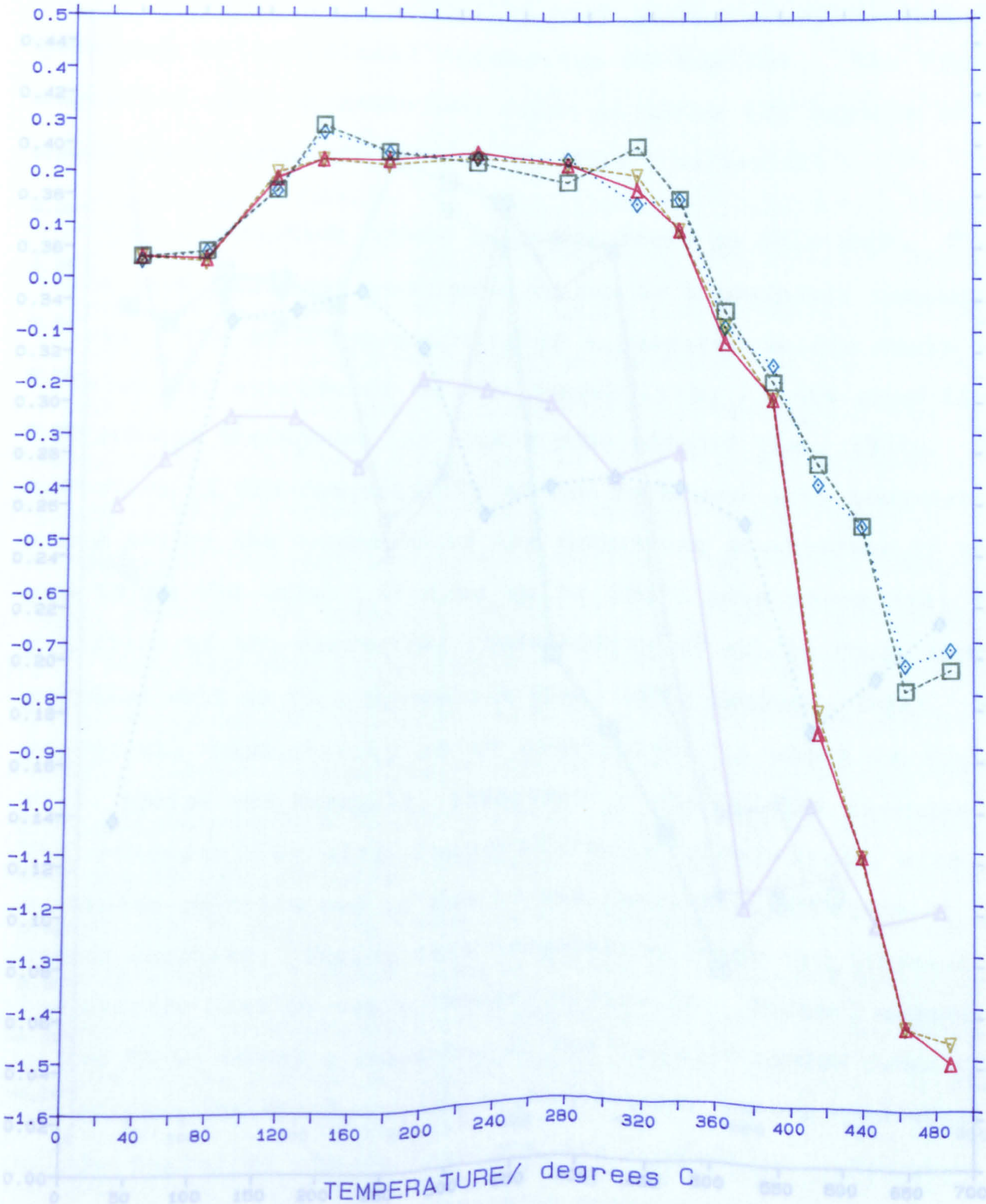


FIG. F9.31. GGRAN1TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PARALLEL TO THE GRAPHITE FOLIATION

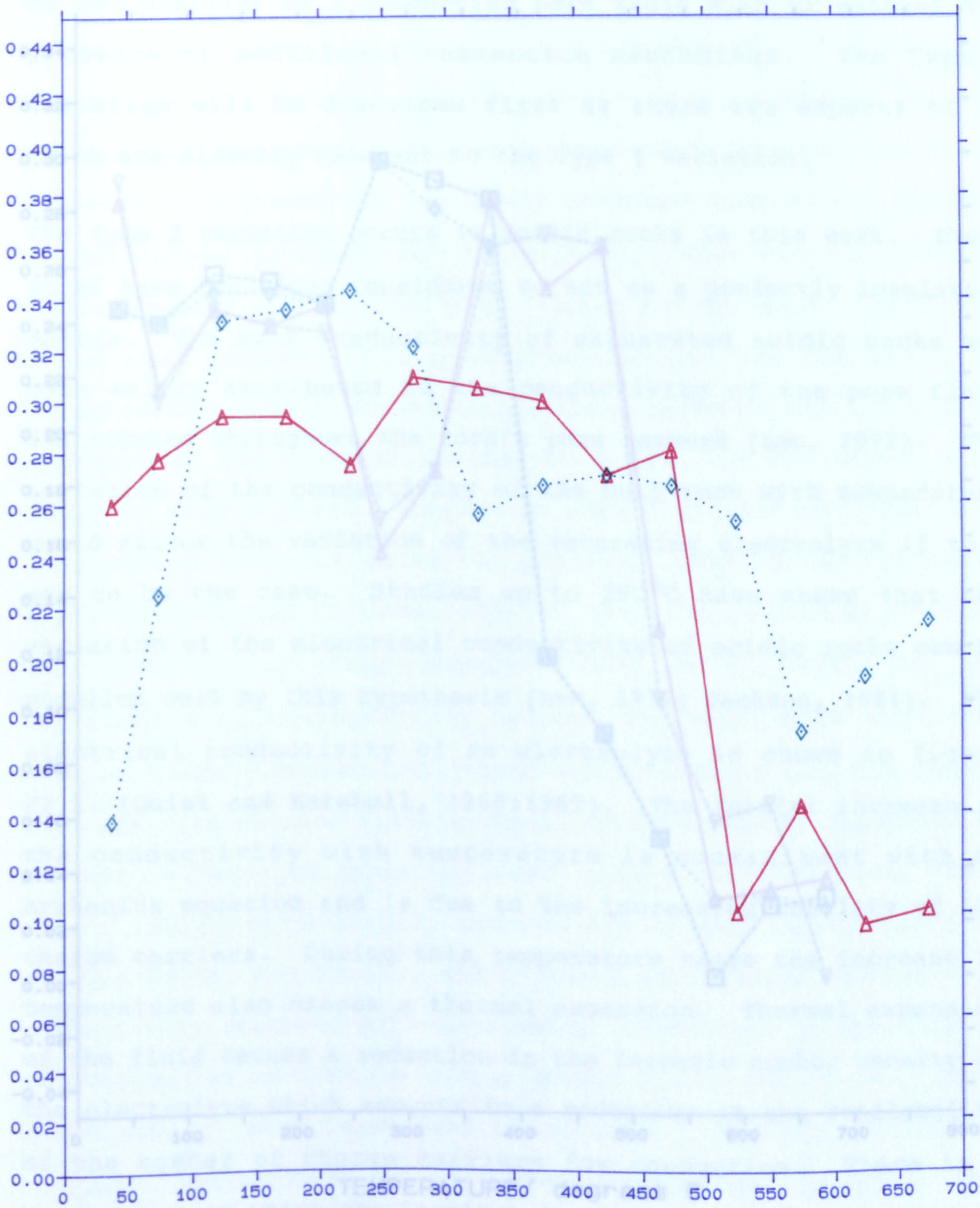
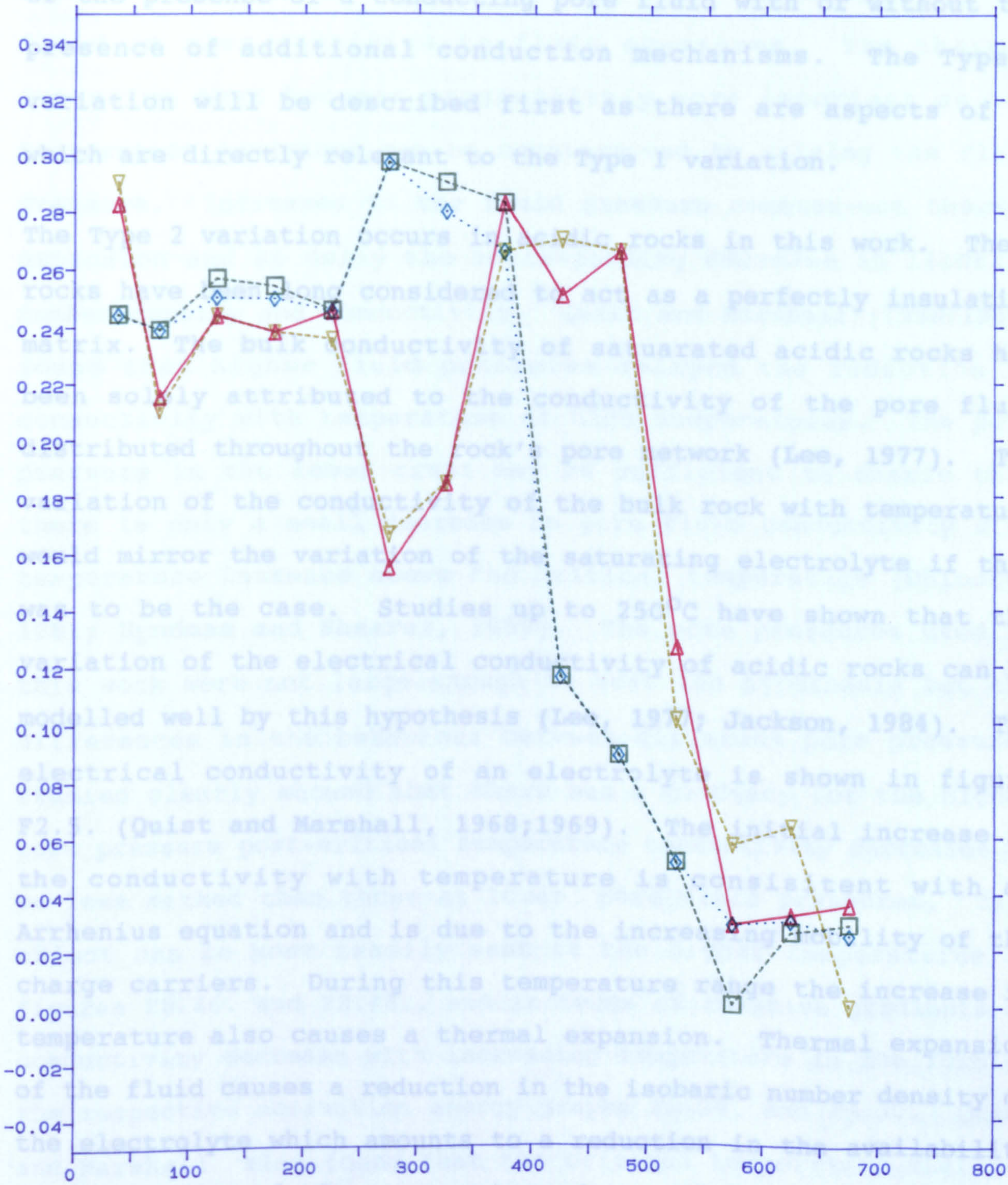


FIG. F9.32. GGRAN2TA ACTIVATION ENERGY VARIATION WITH TEMPERATURE FOR 0.5M NaCl SOLUTION SATURATED GRAPHITE GRANULITE CUT PERPENDICULARLY TO THE GRAPHITE FOLIATION



9.3.6. ACIDIC (TYPE 2) ROCK TEMPERATURE VARIATIONS

Both the Type 1 and Type 2 variations can be explained in terms of the presence of a conducting pore fluid with or without the presence of additional conduction mechanisms. The Type 2 variation will be described first as there are aspects of it which are directly relevant to the Type 1 variation.

The Type 2 variation occurs in acidic rocks in this work. These rocks have been long considered to act as a perfectly insulating matrix. The bulk conductivity of satuarated acidic rocks has been solely attributed to the conductivity of the pore fluid distributed throughout the rock's pore network (Lee, 1977). The variation of the conductivity of the bulk rock with temperature would mirror the variation of the saturating electrolyte if this was to be the case. Studies up to 250^{o}C have shown that the variation of the electrical conductivity of acidic rocks can be modelled well by this hypothesis (Lee, 1977; Jackson, 1984). The electrical conductivity of an electrolyte is shown in figure F2.5. (Quist and Marshall, 1968;1969). The initial increase in the conductivity with temperature is consisitent with an Arrhenius equation and is due to the increasing mobility of the charge carriers. During this temperature range the increase in temperature also causes a thermal expansion. Thermal expansion of the fluid causes a reduction in the isobaric number density of the electrolyte which amounts to a reduction in the availability of the number of charge carriers for conduction. There is a temperature at which the increase in conductivity due to mobility increase is exactly balanced by the reduction in conductivity due to isobaric number density reduction. At this critical

temperature there is a maximum conductivity. The conductivity at post-critical temperatures decreases due to the further increase of the isobaric number density effect at the expense of the effects dominating at low temperatures. The conductivity of the fluid is also sensitive to fluid pressures. The thermal expansion that becomes progressively more important as the temperature increases can be counteracted by raising the fluid pressure. Increases in the fluid pressure counter-act thermal expansion and so delay the corresponding decrease in isobaric number density and conductivity. **Quist and Marshall (1968;1969)** found that higher fluid pressures delayed the reduction in conductivity with temperature at high temperatures. The pore pressure in the lower crust may be sufficient to ensure that there is only a small decrease in pore fluid conductivity with temperature increase above the critical temperature **(Ohloeft, 1981; Hyndman and Shearer, 1989)**. The pore pressures used in this work were not large enough to test the hypothesis but the differences in the behaviour between different pore pressures studied clearly showed that there was a tendency for the higher pore pressure post-critical temperature conductivity decreases to be less marked than those at lower pore fluid pressures. This effect can be most readily seen at the higher temperatures of figures F8.46. and F8.48., and in terms of relative gradients of conductivity decrease with increasing temperature in the form of the respective activation energy graphs F9.28. and F9.30. Quist and Marshall also found that the critical temperature migrated to higher temperatures as the fluid pressure was increased. The effect of fluid pressure upon the conductivity of the electrolyte at pre-critical temperatures was found to be very small due to the relatively low compressibility of the fluid under these

conditions. Figure F2.5. sums up the variation of the electrical conductivity of an electrolyte very well. It would be expected that the temperature variation of saturated rocks would be analogous to that of the saturating electrolyte if the electrical conductivity of a saturated rock was due entirely to the pore fluid saturating its pore spaces.

Figures F8.46. and F8.48. show the rocks undergoing Type 2 variations. It can be seen that the variations mirror the variation of the conductivity of the electrolyte. Several comparisons between the variations can be made:

1. The general shape of the two variations is very similar to that of the pore fluid alone (F2.5.)
2. The increase of the conductivity of the Type 2 variation with temperature at low temperatures and that of the electrolyte are both such that they can be described by one or two activation energies in the pre-critical temperature region.
3. The critical temperature of the Type 2 variation (around 350^{o}C) is very similar to that of the electrolyte (from 325^{o}C to 375^{o}C for the fluid pressure varying from 0.1 to 0.2 GPa).
4. The post-critical temperature conductivity of both the Type 2 variation and the electrolyte variation show large negative gradients. The 'inverse' variations of both the electrolyte and the Type 2 variations show smooth non-linear variations that progress away from the pre-critical linear variations. This would not be expected from a system that obeys an Arrhenius model. The post-

critical temperature variation is due to decreases in the charge carrier availability as has already been said. This is a breakdown in an assumption upon which the 'inverse' variation is constructed and explains the markedly non-linear effect.

5. The variation of both conductivity variations is smooth indicating a gradual change in the process of conduction.

6. The rate of Type 2 conductivity decrease at post-critical temperatures is less at higher pore fluid pressures in a similar way to that which applies to the electrolyte variation.

7. The rate of Type 2 conductivity increase at pre-critical temperatures is slightly greater at higher pore pressures than for lower pore pressures. This corresponds to there being no significant difference in the case of the electrolyte variation. While the small Type 2 effect may be due to the compressibility of the fluid increasing the charge carrier number density slightly, it is more likely to be due to the way the increased pore fluid pressure interacts with the rock matrix increasing the porosity slightly. It is this increase in the amount of fluid present that is more likely to cause the small increase in the rate of conductivity increase at the higher pore pressure.

8. There is no obvious migration in the critical temperature with pore fluid pressure in the Type 2

variation rocks. A small positive migration would be expected from the electrolyte data.

9. The activation energies of the Type 2 variation for differing pore pressures are very similar until around 400^{o}C whereupon they diverge. The post-critical changes in the charge carrier density ensure that the post-critical activation energies have no real physical meaning. These apparent activation energies are calculated from data for conditions under which the Arrhenius equation assumption does not apply. The apparent activation energy is, however, useful to the extent that it gives a quantitative measure of the rate of log conductivity decrease with increasing values of 1000/T(kelvin). The value of the apparent activation energy at these high temperatures clearly shows that the rate of reduction of conductivity with temperature is less for samples held at higher pore pressures as noted in point 6 of this list and seen in the electrolyte conductivity data.

10. The value of pre-critical temperature activation energy for these type 2 rocks was found to be 0.22 eV with a range of 0.02 eV either side (Table T9.7.) compared with a value of 0.2 eV found by **Jackson (1984)** for saturated quartz rich rocks and the values of 0.2 eV found for 'brine' by **Parkhomenkho (1982)** and 0.14 to 0.2 eV depending temperature (from 190 to 0^{o}C) for 0.5M NaCl found by **Waxman and Smits (1968)**(Figure F2.12.).

TABLE T9.7. ACTIVATION ENERGIES FOR SAMPLES AT HIGH PRESSURES AND TEMPERATURES

SAMPLE TYPES OF ROCK	ACTIVATION ENERGY (eV) FOR TEMPERATURE REGIONS V.LOW 25-75°C	LOW 75-350°C	MEDIUM 350-600°C	HIGH 600-900°C
TYPE 1				
BASIC AMPHIBOLITE				
IVD1	0.04	0.25	0.49	0.13
BASIC GRANULITE				
IVE1	-	0.20	0.47	0.20
ORTHOPYROXENITE				
MER21	0.12	0.18	0.38	0.13
MER22	-	0.14	0.32	0.16
AMPHIBOLITE GNEISS				
A8B	0.07	0.25	0.50	0.15
L32	0.15	0.21	0.46	0.05
GRAPHITE GRANULITE				
GGRAN1	-	0.28	0.31	0.12
GGRAN2	-	0.25	0.30	0.04
MEAN	0.095	0.22	0.40	0.12
TYPE 2				
GRANODIORITE				
L33B	0.07	0.22	(-1.3)	(-)
GRANITE				
V1	0.05	0.22	(-1.4)	(-)
MEAN	0.06	0.22	(-1.35)	(-)

Apparent activation energies are given in parentheses.

The vertical shift in the higher pore pressure variations of the figures F8.46. and F8.48. is not comparable to the electrolyte variation. This offset is due to the way the increased pore pressure interacts with rock matrix opening pores and is as such an effect upon the conductivity via a different process than the effect of temperature upon the conductivity of the electrolyte.

The small section of variation at temperatures lower than 100^{o}C in both of the Type 2 variations, and present in some of the Type 1 variations, cannot be explained by the saturation of an insulating matrix with a conductive fluid. The breakdown in the linear portion of the Type 2 variation when plotted to show the inverse variation (figures F8.58. and F8.60.) is due to the assumption of constant charge carrier availability breaking down for reasons given previously. The assumption becomes gradually less applicable as the temperature increases to produce a smooth variation away from a linear variation in the 'inverse' variation plots.

It can be seen that the pore fluid saturation model can describe the variation of Type 2 very well. There are several important conclusions that can be drawn from this observation. Firstly that there is no obvious additional conduction mechanism opperating in these types of rocks over the range of pressures and temperatures used in this work. Secondly that the variation can be seen to describe the variation in conductivity due to a saturating electrolytic fluid in any rock of a comparable pore matrix and porosity providing that there is no additional conduction mechanism and can be considered to be a basal conductivity in those rocks that do posses an extra conduction mechanism. Thirdly and finally, that the decrease in conductivity at the higher temperatures demonstrates that the argument between geophysicists and geochemists about the necessity for conducting electrolytes in the lower part of the lower crust may be largely futile. The experimental results indicate that even if water based electrolytes were present they

would not in themselves contribute much to the conductivity. It should be noted, however, that the pore pressures found in the lower crust are most likely to be higher than those used in this work. Increasing pore pressures has been observed to minimise the post-critical temperature decrease in the electrical conductivity (**Quist and Marshall, 1968; 1969**) and is thought by some authors that the pore fluid pressure in the lower crust may be sufficient to cause there to be only a minimal decrease in the electrical conductivity with increasing temperature after the critical temperature (**Ohloeft, 1981; Hyndman and Shearer, 1989**). The presence of water in the lower part of the lower crust may still be important in activating other mechanisms such as clay counterion conduction. Very little is known about the effect of high temperatures and pressures upon this type of conduction, however there are plans to investigate the way in which counterion exchange capacity changes with pressure and temperature using a gamma ray tracer technique (**Glover, 1989**). It is interesting to note that the maximum conductivity attained by the Type 2 variation occurs at a temperature similar to that of the lower crustal high conductivity layer (**Jones, 1987; Etheridge et al., 1983;1984; Jodicke et al., 1983**). Further geological implications will be reserved until chapter 10.

9.3.7. BASIC (TYPE 1) ROCK TEMPERATURE VARIATIONS

One of the most striking features of the type 1 variation rocks is their mutual similarity considering that there was a range of three orders of magnitude between their low temperature conductivity values, and the range of their variation between two and three orders of magnitude. This feature is probably the single most important pointer to there being a limited number of

processes causing the electrical conduction within saturated rocks. The absence of one or more may cause the dramatic difference in variation observed between type 1 and type 2 rocks as has been discussed previously.

The Type 1 variation must have an explanation that is based upon that of the Type 2 variation. This is because it too refers to rock samples consisting of a matrix of pores that were saturated with an electrolytic pore fluid. If the matrix were the same as that in the acidic rocks then the variation would be of the same shape. The obvious difference between the two variations must be either due to some difference in the rock matrix itself or the way the rock matrix interacts with the pore fluid. It is interesting to note that in all cases of the Type 1 variation (figures F8.39. to F8.45., F8.47. and F8.49. to F8.50.) the shape of the temperature variation up to around 400^{o}C is very similar to that of the Type 2 variation. Up to this point the conductivity variation can be attributed mainly to pore fluid conduction. It has, however, already been shown that the absolute levels of the rock conductivity of type 1 rocks is higher than that of type 2 samples of similar porosity and that this observation leads to the inference that an additional conduction mechanism operates in type 1 rocks. Previously discusssion of figure F9.20. has shown that the shape of allvariations up to 350^{o}C are similar. The graphite granulite samples undergo greater variations with temperature than the other rocks studied in this work. This will be discussed in greater detail later in this section. The majority of type 1 basic rocks also have large variations of conductivity with

increasing temperature, although not as large as the graphite granulites. The acidic (type 2) rocks and the orthopyroxenite showed the least change in conductivity with temperature although the range of conductivities for a temperature range of 0 to 900°C was two orders of magnitude.

The Type 1 and Type 2 variations begin to diverge at temperatures around the critical temperature. The rapid decrease in gradient through zero and becoming largely negative by the Type 2 variation is not seen in the Type 1 variation. There is in fact an increase in the positive gradient; the rate of temperature increase becomes larger. This effect can be explained by the addition of a conduction mechanism other than through the pore fluid. The immediately subsequent variation can be described by the presence of a process with an activation energy higher than that of the initial variation. The larger activation energy must be the result of the summation of the electrolytic process and another process. There is evidence for such a process in the analogous behaviour found in the temperature variation of dry basic rocks by Jackson (1984). Jackson has found that the conductivity of dry basic rocks, low at low temperatures, shows a rapid increase at temperatures exceeding 400°C. At such temperatures the sample would be completely anhydrous. The mechanism by which conduction is taking place can not therefore be clay counterion conduction and is most probably attributable to the intrinsic conduction within mineral lattices at raised temperatures. Such conduction has a temperature variation that can be described by an Arrhenius equation and gives a single value of activation energy. The activation energies reported by Jackson were in the range 0.25 eV to 0.90 eV compared with 0.30

eV to 0.50 eV found in this work.

Another change occurs in the variations at temperatures above 600°C. The variation at temperatures past 600°C varies from sample to sample amongst the basic rocks. Some of the samples show a sudden change of gradient to a considerably lower one. Such samples were the basic amphibolites, IVD1 and IVD2, and the amphibolite gneisses, A8B and L32, (Figures F8.39., F8.40., F8.45. and F8.47.), while some such as the orthopyroxenites, MER21 ansd MER22 and the graphite granulites, GGRAN1 and GGRAN2 show a more gradual change to a lower gradient which may at first sight seem like a continuation of the progressive decrease in log conductivity with temperature produced by Arrhenius variations (Figures F8.43., F8.44., F8.49. and F8.50.). Two samples, the basic granulites, IVE1 and IVE2 show more complex behaviour (Figures F8.41. and F8.42.).

The shapes of these variations are best compared by reference to the 'inverse' temperature variations (Figures F8.51. to F8.57., F8.59. and F8.61. to F8.62.). All the type 1 samples exhibited a low temperature linear portion analogous to that found over the same range in the type 2 variations but they had lower absolute values of conductivity. It has already been said, that this data points towards electrolytic pore fluid conduction as the main mechanism for conduction but with the need for an additional mechanism to be supplementing the type 1 rock conductivity values. A comparison was made between the activation energies that these portions of type 1 and type 2 curves represent from data shown in the activation energy plots (Figures F9.21. to

F9.32.). The average activation energy of the type 1 portions was 0.22 eV with a standard error in the mean of 0.016 eV; the average activation energy of the two type 2 portions was 0.200 eV with a full range of 0.020 eV which has already been shown to agree with the measurements of brine activation energies of other authors. This is strong evidence for both of these low temperature portions of the variation being primarily due to electrolytic conduction. It can be seen that there is a tendency for type 1 rocks to have higher activation energies and therefore greater variations of electrical conductivity with temperature that type 2 rocks although the differentiation between the two groups is not well marked. It may also be seen that the data presented here agrees with that presented in table T9.6. in concluding that the bulk of conduction in both types of rock at low temperatures is dominated by electrolytic pore fluid conduction. Type 1 rocks have a small additional conduction mechanism that does not have great effect upon the bulk rock conductivity at low temperatures but begins to dominate the electrolytic conduction at temperatures above around 350^{o}C in concert with the results found be **Jackson (1984)** during experiments on the electrical conductivity of anhydrous rocks at high temperature.

The most noticeable variation from this agreement of activation energies is exhibited by the graphite granulite samples, GGRAN1 and GGRAN2 (Figures F9.31. and F9.32.). These have activation energies in the low temperature range of 0.28 eV and 0.25 eV respectively. The increase in the activation energies is probably due to the effect of the presence of conduction through the intergranular graphite.

The 'inverse' temperature graphs of the type 1 samples show a linear portion of greater gradient a temperature greater than 350°C than that below 350°C in most cases although the temperature range for which this portion occurs varies (Figures F8.51. to F8.57. and F8.59.). The average activation energy associated with these variations was calculated from data shown in figures F9.21. to F9.27. and F9.29. to be 0.400 eV with a standard error in the mean of 0.03 eV. This part of the variations is believed to indicate the operation of an additional conduction mechanism. It is interesting to note that the activation energy compares with 0.25 to 0.90 eV obtained by **(Jackson, 1984)** from measurements on similar but dry samples over a similar temperature range. The work done by **Jackson (1984)** on dry rocks of a similar type to those in this work showed very low activation energies (average around 0.07 eV) up until 300°C to 400°C whence the activation energies abruptly became much higher (average around 0.66 eV) due to the rapid increase of mineral conduction. It is postulated that the increase in the conductivity of the type 1 rocks at about this temperature is due to the effect of this increase in mineral conduction and that the value of the activation energies found (average 0.40 eV) are not as high as those found by **Jackson (1984)** due to the detractive effect of the rapid reduction in electrolyte conduction. The small activation energies that Jackson found at the low temperatures were probably due to a certain amount of fluid being present in his nominally initially dry samples and would be included in the electrolytic conduction undergone by the saturated rocks in this work.

The last portion of the curve has variable characteristics dependent upon the sample. Those samples showing an obvious break in the temperature variation such as the basic amphibolites, IVD1 and IVD2, and the amphibolite gneisses, A8B and L32 (Figures F8.39., F8.40., F8.45. and F8.47.) show a third reduced-gradient linear portion in the high temperature range of the 'inverse' temperature graphs (Figure F8.51., F8.52., F8.57. and F8.59.). The gradient, and hence activation energy, of these portions seems to be dependent upon the pore pressure. This will be discussed further later in this section. Those temperature variations seeming to have a smooth variation of log conductivity with temperature at higher temperatures such as the orthopyroxenites, MER21 and MER22, and the graphite granulites, GGRAN1 and GGRAN2 (Figures F8.43., F8.44., F8.49. and F8.50.) also show the presence of a third linear portion in the 'inverse' temperature variation and an associated lower activation energy. The difference in the activation energies associated with these graphs is clearly shown in figures F9.21. to F9.32. and the activation energies of all samples are shown in table T9.7. The samples IVE1 and IVE2 (Figures F8.53., F8.54., F8.41. and F8.42.) showed more structure within the upper temperature region for which no explanation is known. Indeed the reason for the sudden reduction in activation energies from around 0.4 eV to an average of 0.12 eV is difficult to explain. It may be due to one or more of the following:

1. Phase changes in the minerals.
2. Chemical alteration of the minerals.
3. The reduction in overall conductivity due to further drop in electrolyte conduction; a drop

beyond which the alternative conduction mechanism begins to fail to compensate for.

Phase changes are unlikely at the pressures that these samples were subjected to. The last two reasons are more likely though it must be said that any increase in alteration might be thought to increase the amount of ions dissolved in the pore fluid which would tend to increase the apparent activation energy.

Finally it should be noticed that the gradients and hence activation energies vary depending upon pore pressure; the lower the pore pressure the lower the activation energy. This effect is small at low temperatures and becomes significant at temperatures above 350^{o}C such that there is a significant difference at the temperatures between 600 and 850^{o}C (Figures F8.45. and F9.27. illustrate the point). It is thought that this behaviour is as a result of the reduction in electrolyte conduction dependency upon pore fluid pressure described in the discussion on type 2 samples. The greater high temperature activation energy for high pore pressures would result due to the smaller reduction in the electrolyte conduction. This mechanism would explain the lack of divergence between the activation energies until after the critical temperature (around 350^{o}C) and would also seem to indicate that the reduction in the gradient and activation energy at the high temperatures is due to item 3 in the list in the previous paragraph.

9.3.8. THE TEMPERATURE VARIATION OF GRAPHITE GRANULITE SAMPLES

The variation of the graphite granulite samples require special mention as they may be directly representative of rocks

dominating the lower crust (Shankland, 1989). Laboratory measurements of the electrical conductivity of graphite bearing basic rock samples at high pressure and temperature are especially important in respect of recent interrest in the use of graphite grain boundary depositions to explain highly conducting horizons within the lower crust (Frost et al., 1989) and recognition that graphite may be present in more lower crustal rocks than was once thought (Jackson et al., 1988; Mathez and Delaney, 1981; Sternberg, 1979). Indeed Frost et al. (1989) have recognised that the presence of graphite might be significant in lower crustal rocks of the following three types;

1. pelitic schists and gneisses
2. mafic plutonic rocks
3. charnockitic and granulitic rocks

It may be that a very thin, grain covering, film of graphite as seen by Frost et al. may be able to produce higher overall conductivities per volume of graphite than graphite present in the form of coarse grains.

The graphite encountered in the four rock samples studied in this work, GGRAN1-4, is in the form of coarse elongate foliated grains. There may have been thin films of carbon coating grain surfaces in addition to the obvious graphite but the techniques used by Frost et al. (1989) were not available to check this.

The mechanisms by which graphite can be deposited in crustal rocks is of great importance to any argument concerning the possibility that graphite might be important to the electrical behaviour of the lower crust. Several mechanisms will be discussed. It is interesting to note that these methods have

important connections with the formation of granulite facies rocks rich in magnetite and ilmenite.

It has been noted that some parts of the lower continental crust contain high proportions of carbon dioxide that originates in the mantle (**Touret, 1971**). Fluxing of the lower continental crustal rocks by mantle derived CO_2 (**Collerson & Freyer, 1978; Jackson et al., 1988; Harris, 1989**) has been cited as a prime factor controlling the formation of granulite facies rocks by forcing dehydration reactions to take place. The replacement of aqueous pore fluids by CO_2 can upset the local dehydrational equilibrium. Keeping the same pressures and temperatures, the removal of the aqueous phase causes more dehydrational reactions to take place producing granulite facies mineralogies and free grain boundary water. Further fluxing may remove this water causing the reactions to continue leaving the pores of the rock full of a CO_2-rich fluid. The onset of granulite facies metamorphism by this method may be controlled by the presence or lack of CO_2 (**Janardhan et al., 1979; Jackson et al., 1988**). Such an occurrence has been noted in granulite formations of Southern India (**Janardhan et al., 1979; Harris, 1989**). Thus fluxing by CO_2 can 'produce' water at depth as well as a granulite mineralogy. The presence of CO_2 rich pore fluids cannot explain the high conductivity layers observed in itself (see chapter 3) but the CO_2-rich fluid can be the source for carbon deposition. Grain boundary graphite may be deposited on cooling by the reduction of a CO_2-rich pore fluid or by the process of oxyexsolution by which titanomagnetite clears itself of titanium (**Fuhrman et al., 1988**).

$$6Fe_2TiO_4 + CO_2 \rightleftharpoons 2FE_3O_4 + 6FeTiO_3 + C \qquad E9.17$$

in magnetite | in pore fluid | magnetite | ilmenite | graphite

The presence of magnetite, titanomagnetite and ilmenite in most of the basic rocks studies and in particular in the granulite facies rocks indicates that oxyexsolution may be a valid mechanism in the rock samples in this work. It can be seen that the presence of a granulite mineralogy, titanomagnetite, magnetite, ilmenite and graphite rock may all be due to the presence of carbon dioxide and that the carbon and water produced by these processes may be able to add to the electrical conductivity of the lower continental crust. It should also be mentioned that granulite facies metamorphism can take place by other mechanisms such as the metamorphism of already anhydrous mineralogies (**Lamb and Valley, 1984**) and the extraction of hygroscopic granite melts (**Nesbitt, 1980**). Recent work (**Peterson and Newton, 1989**) has reported that the presence of a CO_2 rich fluid can depress the melting point of certain rocks by as much as 50^{o}C and promote partial melting. This finding concurs with the findings that the lower crustal rocks are depleted in magmaphile elements. Depletion of elements which preferentially partition into a melt leads to the inference that magma has been extracted from such rocks. Indeed the fact that water will partition into a coexisting melt phase has been used as a possible mechanism for dehydrational metamorphism. The temperatures needed for partial melting of the lower crust are not found in any but the most active tectonic areas. The presence of CO_2 may promote melting in areas where the temperatures are sufficiently high but the temperatures and the

presence of CO_2 indicates that dehydrational metamorphism would have already occurred at shallower layers (Harris, 1989). The presence of CO_2 may also affect the presence or absence of partial melts in the lower part of the lower crust.

Frost et al. (1989) noted that graphitisation in the form of thin grain boundary films may suffer from irreversible disconnection and associated permanent loss of conductivity when decompressionally strained. It was for this reason that they doubt whether laboratory measurements could reproduce lower crustal graphite rock high conductivities. It has already been shown in section 9.1 that reconnection upon increase of confining pressure does occur leading to increases in electrical conductivity at high confining pressures. The temperature variation of the electrical conductivity of graphite granulites is also of interest as it too, differs from the other metabasic rocks studies.

Reference to figure F9.20 shows that;

1. The variation of the electrical conductivity of the graphite granulite samples is greater than the other metabasic rocks studied.
2. The graphile foliation anisotropy present in the samples effects the way the bulk rock conductivity varies with temperature in a similar way to that for the variation of conductivity with pressure. The sample cut parallel to the graphite foliations (GGRAN1) had a greater conductivity variation than

that cut perpendicular to the graphite foliations (GGRAN2).

It can be postulated that these points are due to the presence of graphite conduction in addition to the electrolyte conduction. The size of the increase in conductivity of a rock will depend strongly upon the amount and distribution of the graphite in the rock.

The activation energy of the samples GGRAN1 and GGRAN2 during the higher temperature ranges did not show any sudden changes. GGRAN1 (Figures F8.61. and F9.31.) does not show an increase in the rate of change of log conductivity with 1000/T(kelvin) and hence has a reasonably stable activation energy. Such an effect would be expected as no threshold temperature needs to be reached before graphite mineral conduction can occur as is the case with mineral semiconduction in hydrous and magnetite minerals. Graphite conduction occurs in these rocks as an additional conduction mechanism throughout the whole range of temperatures used in these experiments. The minimum temperature for electronic conduction does not occur for graphite conduction due to channelling of shared valance electrons along planes in the graphite lattice. GGRAN2 (Figures F8.62. and F9.32.) shows a small increase in the activation energy from 0.25 to 0.28 eV but this difference is virtually indistinguishable in the 'inverse' temperature variation graph (Figure F8.62.). Any small changes in Eo may be due to associated mineral conduction in addition to the graphite conduction already present. Minerals such as hydrous amphiboles or magnetite-like minerals, if present, make additional conduction contributions when the temperature is sufficiently high. It is thought that any increases in the Eo

value of these samples due to graphite conduction may also mask the transition from a electrolytic and graphite conduction regime to a mineral conduction and graphite conduction regime.

Another possibly important factor to consider is the formation of intercalated graphite compounds. Graphite is capable of forming lamellar compounds with a variety of elements in which the bonding is primarily ionic. Positive ions form layers between those of carbon atoms. Intercalation can occur between every carbon layer, every second, third, etc. The formation of such compounds is not easy especially for lithium and sodium but is progressively simpler for higher periods and group elements. Taking sodium as an example; sodium ionises and donates an electron to the delocalised sp^2 boding orbital cloud between the layers of graphite (Muller and Kleiner, 1962).

$$Na \longrightarrow Na^+ + e^-$$

e^- → donated to sp^2 cloud

The donation of such electrons increases the charge carrier (electron) density and thus increases the conductivity. The presence of such compounds in the Earth's crust is purely hypothetical. The laboratory synthesis of $C_{64}Na$ requires temperatures of 450°C molten sodium and a long time scale. The product is a deep violet solid with a sodium layer every eighth carbon interlayer space. There may be a mechanism by which high temperatures and long time scales can convert NaCl in aqueous solution and graphite to similar compounds. One thing is certain, the presence of such highly conducting compounds would greatly increase the bulk conductivity of lower crustal rocks

even if only present in small amounts.

The conductivity increases with increasing metal ion/carbon ratio as more electrons are donated into the conducting delocalised sp^2 clouds. The conductivity of C_8K at 77K is 50 times that of graphite (half that of copper) (Bailar et al., 1973). As temperature increases, the conductivity of graphite increases as the charge carriers become more mobile. There is, however, a decrease in the conductivity of the lamellar C_8K compound. If this trend continues then the conductivity advantage given by these compounds may be irrelevant. There is a growing realisation that the conductivity increase due to intercalation of graphite by alkali metal ions is about 10 times along the basal planes and about 320 times perpendicular to the basal layering of the graphite crystallites. This marked anisotropy may have relevance to the way such materials might affect the conductivity of rocks if present (Fischer, 1979).

In summation, the presence of graphite and CO_2 in the lower crust may be able to explain complex mechanisms operating in those areas. Specifically it may be able to explain the lower crustal high conductivity layers and the granulite mineralogy of the lower crust.

9.4. THE EFFECT OF CONFINING PRESSURE UPON THE ELECTRICAL CONDUCTIVITY OF SATURATED ROCK SAMPLES AT HIGH TEMPERATURES AND PORE-PRESSURES

The variation of the electrical conductivity with confining pressure at constant raised temperatures as high as those used in this work has not, hitherto, been done. Up until recently the most comprehensive work of this type was done by Lee (1977), Evans (1980) and Jackson (1984). They were able to measure the confining pressure response of electrical conductivity at temperatures up to 250°C. Higher temperatures were not possible due to the thermal breakdown of the sample sleeve. Measurements have been made up to 900 °C in this work.

The variation of the electrical conductivity with pressure is an important function to measure accurately as it is upon this data that extrapolations are made into the deep crust. Such extrapolations have been necessary because of the lack of high pressure and high temperature electrical conductivity data.

The data obtained in this work have been displayed in two ways;

1. Parametrically (Figures F8.63. to F8.69.)
2. In three dimensions (Figures F8.70. to F8.77.)

Measurements of the variation of electrical conductivity with confining pressure at raised temperatures and pore pressures were done on the samples of orthopyroxenite (MER21), basic amphibolite gneisses (A8B and L32), granite (V1), granodiorite (L33B) and graphite granulites (GGRAN1 and GGRAN2). The electrical conductivity decreased with application of confining pressure in

all cases. The size of the decrease per unit confining pressure became less as confining pressure increased and as temperature increased as shown in figures F8.63. to F8.69. The variation of electrical conductivity with confining pressure had become negligible at temperatures above 600 ^{o}C. The size of the confining pressure variation has direct relevance to geophysical studies of the lower continental crust. One of the traditional methods for obtaining values of lower crustal conductivity has been to extrapolate measurements made in laboratories. Such measurements as these have subjected rocks to lower crustal pressures but never has this been done at lower crustal temperatures. Measurements used in extrapolations have been similar to those pre-transient measurements at around 25 ^{o}C shown in figures F8.1. to F8.12. Comparison of the size of this variation with that of those measured at raised temperatures (Figures F8.63. to F8.69.) clearly shows that too great a variation has been assumed to take place. The extrapolations have overestimated the effect of the confining pressure variation, producing an underestimation of the predicted lower crustal electrical conductivity.

The variations for sample A8B is taken as an example (Figures F8.7. and F8.64.):

1. The variations in electrical conductivity at 25^{o}C (laboratory conditions) for a variation of 0.1 GPa in confining pressure is : -58%
2. The variation in electrical conductivity at 400^{o}C for a variation of 0.1 GPa in confining pressure is : -34%

3. The variation in electrical conductivity at 700^{o}C for a variation of 0.1 GPa in confining pressure is : -19%

4. The variation in electrical conductivity at 900^{o}C for a variation of 0.1 GPa in confining pressure is : -11%

A similar scenario applies to all the other samples studied in this way as can be seen by comparison of figure F8.64. with figures F8.63. and F8.65. to F8.69.

There are several explanations for the decrease in the sensitivity of electrical conductivity to changes in confining pressure as temperature and confining pressure increase:

1. It has been suggested that the presence of enhanced conduction mechanisms may have been compensating for the reduction of electrical conductivity with confining pressure at the higher temperatures. If this is the case then the enhanced conduction mechanism must increase with confining pressure as well as with temperature. The increase with confining pressure should also be proportionately greater at the higher temperatures. There may be a such a mechanism in the case of rocks with high proportion of conductive minerals. A good example of this type of rock is the graphite garnet granulite measured in this work. The confining pressure variations of this rock are shown in figure F8.11. to F8.14.

and F8.68. to F8.69. The electrical conductivity of this rock increases with confining pressure when dry and has a reduced decrease with confining pressure when fully saturated. There was a noticeable increase in the dry electrical conductivity with confining pressure at a raised temperature. This rock sample is unusual in containing a high proportion of a conductive mineral and will be dealt with in greater detail later in the section. Although this may explain the electrical conductivity variation in this one type of rock it is not likely to be applicable to the other rocks studied in which reduced confining pressure variation at high temperatures was also found (Figures F8.63. to F8.67.). Most importantly, there were very similar results for the acidic rocks (Figures F8.65. and F8.67.), rocks for which all other data suggests the absence of all enhanced conduction. It is therefore possible to say that, although there seems to be an enhanced conduction mechanism to explain the observations, and that this mechanism may be important in the case of a few types of rocks, the most likely mechanism for all rocks is likely to be found elsewhere.

2. It should be remembered that the decrease in electrical conductivity of saturated rocks upon increasing confining pressure is due to the loss of pore space in which a conductive fluid can

exist. Decreases in the bulk rock conductivity can be due to decreases in the conductivity of the bulk fluid saturating the rock. The decreases in bulk pore fluid conductivity at temperatures greater than 350°C would reduce the variation of bulk rock conductivity with confining pressure as well as reducing the overall bulk conductivity of the rock. Compaction of the rock to the same extent leads to lower confining pressure variations of electrical conductivity. This mechanism would hold for acidic rocks. It would also hold for basic rocks with the difference that the confining pressure variation, being pore-fluid conductivity based, would be reduced only to the extent of the pore-fluid conductivity, any conduction based upon other conduction mechanisms would remain. At high temperatures the conductivity of the basic rocks is high due to other conduction mechanisms, the conduction due to pore-fluids is reduced to very minor levels and the variation of electrical conductivity with confining pressure, being entirely due to the conduction through pore-fluids, is reduced.

3. The decrease in the conductivity/pressure variations up to 300°C is not explained by either of the two points above. This indicates that some other mechanism or feedback is operating to reduce the conductivity variation. This may be the thermal expansion of pore fluid keeping pores open and reducing the decrease in pore volume upon

increases in confining pressure.

It has already been mentioned that the electrical conductivity of a saturated rock sample is a much more sensitive function of temperature than of confining pressure. This is well demonstrated by the 3D representation of the electricalconductivity measurements made at high temperatures and high confining pressures and pore pressures. The data is shown in the 3D form in figures F8.70. to F8.77.. Four samples have been shown in this way. These are the acidic granite + granodiorite samples (V1 and L33b respectively) and the basic amphibolite gneisses (A8B and L32). There are several major points to be noted from these diagrams:

1. The variation with the confining pressure is insignificant compared with the variation with temperature.
2. The confining pressure variation becomes less as temperature increases in the case of both the acidic rocks and the basic rocks.
3. The difference in shape between the $\sigma - P_c - T$ surfaces of the acidic and basic rocks is large (compare figures F8.70. and F8.71. with figures F8.72. to F8.77.). The lack of enhanced conduction in the acidic rocks constrains the acidic rock electrical conductivity curves (figures F8.70. and F8.71.) to reach a maximum and then decline in order to be consistent with the behaviour of the electrical conductivity of the pore fluid as described by Quist and Marshall (1968; 1969). The effect, combined with the

reduced confining pressure reduction as temperatures increase produces the saddle-like 3D image shown in figures F8.72. to F8.77. There are three views of each of these σ - P_C - T surfaces in order to give a comprehensive view of all the faces of the 'saddle' which is not possible with one view. The more basic rocks (figures F8.70. and F8.71.) have a σ - P_C - T surface that is different from the saddle-like acidic rock σ - P_C - T surface in that electrical conductivity continues increasing with temperature while the confining pressure varies in a similar way to that in acidic rocks. The way that the electrical conductivity of the rocks has been discussed in detail previously. The 3D representation is a method of display that gives a more comprehensive view of the greatest amount of relevant data.

4. The 3D data sets presented in this work enable a limited geotherm/geobar path to be traced across their surfaces. The tracing of such a path of electrical conductivity with a realistic geotherm/geobar has the ability to convert depth to conductivity without the need to bear in mind that;

$$\sigma = f(P_C, P_p, T)$$

and that P_C and T are not independent.

The pressure constraints on this work have ensured that the mapped space stops at P_C = 0.2 GPa. This is unfortunate as a variation of confining pressure up to lower crustal values would

make the σ - P_c - T a very useful diagram. An approximation to the true variation can be obtained by extrapolating the diagram to higher confining pressures. The extrapolation is most likely to be not significantly in error. This judgement has been made in the light of what is known about the smoothly varying way the confining pressure changes up to 0.4 GPa at temperatures up to 250^{o}C and up to greater pressures at ambient laboratory temperatures in other works listed in table T2.5. Use of one of these diagrams and an assumed geobar/geotherm can produce a crustal electrical conductivity cross-section, such a cross section will be of very little value as it assumes that the crust consists of the one lithology represented in the diagram used. A set or combination of diagrams for several dominant lithologies is needed to provide a meaningful cross-section of crustal electrical conductivity. Knowledge of the dominant lithologies at various depths from studies of xenolith suites and upper and lower crustal exposures can be used to combine σ - P_c - T surfaces for various lithologies. The positions of the horizons between these lithologies in σ - P_c - T space should be constrainable by knowledge of P - T stability fields for the various mineralogies. The choice of a geotherm/geobar would enable a multi-lithology electrical conductivity cross-section to be produced. The assumptions made in such a procedure would inevitably produce large errors. Unfortunately such errors will remain until more work has been done on the measurement of electrical conductivity of saturated rocks at high temperatures and pressures and until deep drilling enables more information to become available concerning deep crustal values of;

1. Saturation
2. Brine salinity

3. Mineralogical stability

4. Composition

5. Temperature

6. Pressure

7. Formation factor

The availability of such data, although of only indirect use to the electrical characterisation of the lower crust, would enable that of the upper crust to be improved. This would enable the study of the lower crustal values to be done with less ambiguity from better known 'boundary conditions'.

CHAPTER 10: CONCLUSIONS

10.1. FULFILMENT OF AIMS

The aim of this laboratory-based experimental study has been to simulate conditions within the Earth's lower crust and to measure the electrical conductivity of certain rocks under these conditions.

The conditions that needed to be satisfied in order to be able to create a lower crustal environment for electrical conductivity measurements included;

- Full rock saturation
- Independent variability of pore pressure up to confining pressure values.
- Confining pressures up to 1 GPa.
- Temperatures up to 900^{o}C.

Up until now most researchers have confined their experiments to those which vary only pressure or temperature. When both are accurately controlled it has tended to be only to the relatively low pressures and temperatures appropriate to upper crustal rocks. High temperature measurements have always been done on dry rocks. The reasons for this have been given in previous chapters. It is sufficient to mention that the measurement of saturated rocks in which the pore pressure could be varied independently of the confining pressure was necessary. The confining pressure, pore pressure and temperature maxima also needed to be increased to encompass those conditions found in the lower crust. At the beginning of the project it was thought that these constraints would generate experimental

specifications that would be very difficult or impossible to meet with the contemporary technology.

Lee (1977) satisfied some of the above conditions in a cell designed to operate in an oil-filled hydrostatic pressure vessel. He was able to vary the pore pressure and the confining pressure up to 0.4 GPa independently and operate at a temperature of 250^{o}C. The main part of this work has been to extend the temperature and pressure range of the measurement cell (Chapter 6). Although the new cell has been designed to operate up to 1 GPa, tests have been possible only up to 0.4 GPa and experiments up to 0.2 GPa due to the lack of suitable pressure vessels.

The increase in temperature required a complete re-design of the experimental cell and the experimental practice. The following main changes were made;

- Design and construction of a new type of cell for operation at 1 GPa and 900^{o}C.
- Exploration of the concept, design, building and testing of a guard-ring electrical conductivity measurement technique to compensate for the heavy leakage currents caused by metal sample jacketing.
- Use of an internally heated gas pressure vessel.
- Design and construction of ancilliary systems for measurements in a gas pressure vessel including the building of furnaces, pore pressure raising systems and control and measurement instrumentation.

The new systems were tested, compared with measurements made in the Lee Cell and calibrated to give conductivity measurements accurate to within an error of 4%.

Measurements of rock conductivity were made upon saturated rocks;

- on cycling confining pressures to 0.2 GPa
- on varying pore pressure to 0.18 GPa
- on cycling temperature to 900°C at various values of pore pressure.
- on cycling confining pressure from 0.1 GPa to 0.2 GPa at a pore pressure of 0.08 GPa and various values of temperature up to 900°C.

Measurements were taken on slow cooling and after full equilibration of temperature and pressure, on initial application of temperature and pressure and after pressure and temperature transients had been removed by cycling or maintenance of temperatures and pressures at a steady level for a considerable period of time.

Inability to experiment up to the planned pressures of 1 GPa arose from the lack of a pressure vessel of sufficient specification. The original plan, to have such a vessel designed and built while the new cell development was being done proved to be impractical in the financial and political climate in which this work was completed. Experiments were done in a large internally heated 0.2 GPa gas pressure vessel situated in the School of Physics in the University of East Anglia. The

vessel was able to provide sufficient pressure to keep the pore fluid **dense** at high temperatures while enabling some pressure variation to be done. Measurement of electrical conductivity as a function of confining pressure at lower crustal high temperatures was also enabled. Measurement of electrical conductivity along a geotherm was not possible due to the restriction on the range of the pressure variable.

Extensive conclusions can be drawn from the data despite the pressure variable restriction. This is because;

- The sensitivity of the rock conductivity to pressure changes is orders of magnitude less than the sensitivity to temperature changes.
- The pressure variation of rock electrical conductivity is reasonably well understood and can be extrapolated with a certain degree of confidence to larger values of pressure. The absolute magnitude of the pressure variations has, however, been shown by this work to be smaller than was previously thought at high temperatures.
- There are many measurements of the electrical properties of saturated and dry rocks at high pressure and either no or relatively low temperatures, but none at lower crustal temperatures. Measurements of rock conductivity as a function of pressure for the high temperatures thought to be found in the lower continental crust have been possible in this work.

Such measurements have hitherto been avoided by experimentalists due to the experimental complexity involved. The results presented here are thought to be unique.

The crux of this work has been the development of a method of measuring the electrical conductivity of saturated rocks at high temperature thus enabling the effect of lower crustal temperatures upon the electrical conductivity of rocks to be observed. The measurement cell has been designed to withstand pressures up to 1 GPa in the event of a pressure vessel of a sufficiently high pressure specifications becoming available.

10.2. SUMMARY OF RESULTS

The majority of the rocks studied were those thought to be characteristic of the lower continental crust. Samples of all the rock types were saturated with 0.5 M NaCl (approximately the concentration of salt in seawater). Some of the rock samples were saturated with double distilled water and some rocks were left unsaturated.

There is some argument whether conducting fluids, specifically water, exist in the pores of lower crustal rocks (**Shankland, 1989**). The measurements of rock samples saturated with 0.5 M NaCl have been considered to set an approximate upper limit to the possible electrical conductivity of lower crustal rocks. There is, however, some evidence for pore fluid of greater molarity existing to considerable depths in the continental

crust from the Kola borehole (Kozlovsky, 1984; 1987). This may increase the upper limit of lower crustal bulk saturated rock conductivity. The observed high concentrations of dissolved solids in deep boreholes may be accounted for by the progressive solution of rock matrix ions into the pore fluid as temperature and depth increases. The lower limit to the electrical conductivity of the rocks can be found from measurements upon dry rocks made by many authors, for example Lastovickova (1984) and her colleagues, and Jackson (1984).

The rocks studied were;

- granite
- granodiorite
- amphibolitic gneiss
- basic amphibolites
- basic granulites
- pyroxenite

The samples could be divided into two groups on the basis of the variation of their electrical conductivity with temperature and pressure as found by Lee (1977), Evans (1980) and Jackson (1984). The two groups are;

- TYPE 1: Low quartz content metamorphic and igneous rocks.
  Low porosity, low quartz, high mafic mineral content and a significantly greater proportion of magnetite and titanomagnetite compared with Type 2.

- TYPE 2: High quartz content rocks.

  Low porosity, high quartz, acidic, igneous or metamorphic rocks with a low proportion of mafic minerals and an especially low occurrence of magnetite and titanomagnetite.

Conduction in Type 2 rocks is accounted for by suggesting that it is wholly due to electrolytic conduction through a pore fluid saturating the rock pores. Conduction by such a mechanism decreases upon the application of effective pressure to a rock sample as pore conduction pathways are progressively cut off by the collapse of cracks and pores. The conductivity of type 2 rocks initially increases upon an increase in temperature as the conductivity of the pore fluid increases. This increase occurs up to about 350^{o}C. Above about this temperature there is a decrease in the conductivity of the saturated bulk rock. This decrease can be ascribed to a decrease in the conductivity of the pore fluid. Such a decrease in the pore fluid conductivity can be explained by isobaric volume expansion of the pore fluid reducing the charge carrier density. The operation of this effect in pure electrolytes has been observed and described by **Quist and Marshall (1968; 1969).** The results reported here include the first observation of a decrease in the bulk electrical conductivity of a saturated rock with an increase in temperature as conductivity measurements of saturated rocks at these high temperatures have not, hitherto, been done. It must be said, however, that the high temperature decreases in electrical conductivity seen in these rocks is reduced at high pore pressures due to the fluid pressure tending to offset any

fluid thermal expansion. The effect of pore pressures has been noted in this work but it is the low values of pore pressure have resulted in differential rates of decrease in electrical conductivity at high temperatures, the decreases are much reduced at high pore pressures. The pore pressure at depth may be sufficiently high to make the high temperature decrease in the electrolytic conduction minimal **(Hyndman and Shearer, 1989)**.

As a first approximation it may be said that the quartz rich rocks of Type 2 behave as non-conducting rock matrices through which pore fluids pass at lower crustal temperatures and that their conductivity can be described solely by considering the effect of temperature upon the saturating pore fluid. Although there is likely to be a certain proportion of overall conductance provided by solid state mineral conduction, especially at the higher temperatures, this is seen to be still a small effect compared with the conductance of the pore fluid. The results described here are entirely consistent with the results of **Quist and Marshall (1968; 1969)**. It has been satisfying to find analogous behaviour in the variation of the electrical conductivity of saturated rocks with low conductivity matrices.

The conductivity of Type 1 rocks can best be described by three regimes. Firstly a low temperature regime in which pore fluid conductance is dominant with a small amount of conduction by other mechanisms. Such mechanisms can be seen to be present from the values of observed electrical conductivity higher than those predictable from knowledge of the rock porosity and the pore fluid bulk conductivity, and the higher rates of change of conductivity with temperature found in type 1 rocks. Secondly a

middle temperature region in which the non-pore fluid contribution increases and compensates for the steady decrease in the rate of electrical conductivity increase with temperature. Lastly a high temperature regime in which the pore fluid contribution decreases to a very small value and the other mechanisms of conduction provide a correspondingly greater proportion of the total conductance. The increase in the contribution from non-pore fluid mechanisms at this level is insufficient to completely compensate for the drop in pore fluid conduction. The pore fluid conductance may not decrease as fast at higher temperatures as it did in the case of Type 2 rocks as the presence of clay minerals lining the pores of the Type 1 rocks may desorb some of the ions that they had previously adsorbed to compensate for the progressive lowering of ionic number density that is due to the isobaric thermal expansion of the pore fluid.

The non-pore fluid component of conduction may be attributable to hydrous minerals and clay alteration products lining the pores and along grain boundaries which may become pores on the application of pore pressure **Jackson (1984)**. It may also be attributable to the presence of spinel minerals especially titanomagnetites. Other relatively more conductive phases, such as graphite, may also play a part in raising the bulk conductivity of the whole rock.

The observation that enhanced conduction was found in rocks that contained a high proportion of Fe^{2+}, Fe^{3+}, Na^{+} and K^{+} compared with Al^{3+}, Mg^{2+}, Si^{4+} and Ca^{2+} confirmed findings made by **Jackson**

(1984) and suggestions made by **Parkhomenko (1982)**. The finding that minerals containing a high proportion of larger ions tend to have greater conductivities, especially at low temperatures leads to the proposition that such minerals as magnetite, titanomagnetite, ferrous garnet and pyroxenes, and sodic and potassic hornblendes may have significant control of the electrical conductivity of the lower crust. All the above mentioned minerals are found in greater proportion in those basic rocks that are thought to be present in the lower crust relative to the more acidic rocks making up the upper crust. At lower crustal temperatures the electronic conduction from this source becomes very significant in the absence or otherwise of a fluid conducting phase **(Parkhomenko, 1982; Lebedev and Shepel, 1981)**. If the rock is dry, as may very well be the case in the lower part of the lower crust, the conductivity from the electronic mineral conduction will be the major mechanism of conduction and hence control the overall conductivity of that part of the crust. The electronic mineral conduction mechanism could still be a significant contributor to the overall rock conductivity even when the rock is saturated, accounting for a large proportion of the overall conductivity, if the temperature is sufficiently high to reduce the electrolyte conductivity.

The presence or absence of clay minerals may also have an effect upon the variation of the electrical conductivity. It has previously been noted that a certain amount of retrograde metamorphism of basic rocks of deep genesis may result in the formation of 'rims' of clay minerals. The presence of tectonic forces can compress and thicken areas of crust. The upper crust of such areas are uplifted and subsequently eroded exposing rocks

of a deep crustal type. Rocks exposed to such a scenario are subject to complex changes of stress and temperature fields and may show extensive deformation. It is recognised, however, that the rocks of deeper genesis are subjected to reducing pressure and temperature as the overburden is eroded away. Such rocks may undergo a retrograde metamorphism. Such a process is very slow, but so is the progressive exposure. 'Rims' of alteration products formed by retrograde metamorphism may be seen in some basic rocks (GGRAN1, GGRAN2, GGRAN3, GGRAN3, PLATES P4.1., P4.8.). Clay-mineral alteration products lining pores have been invoked to explain some of the differences between the pressure variation of the electrical conductivity of basic and acidic rocks of continental origin. A model has been proposed that requires the presence of clay-lined pores to restrict the decrease of conductivity of basic rocks with increases in pressure relative to that of more acidic rocks (Jackson, 1984). The mechanism proposed by Jackson (1984) may be causing some of the enhanced conduction observed in the results from the more basic rocks (Figs, F8.39. to F8.45. and F8.47.) compared with the more acidic rocks (Figs. F8.46. and F8.48.). The continuing electrical patency of otherwise mechanically closed pores at high pressures due to the lining of the pores with hydrous alteration minerals, such as clays, seems to be a major mechanism altering the variation of the electrical conductivity with pressure and temperature. The continuing increase in the electrical conductivity of saturated basic rocks as temperature increases, above the temperature which the electrical conductivity has begun to decrease in acidic rocks, strongly indicates that one or more major replacement conduction mechanisms operate. Jackson's

enhanced clay conduction mechanism may be applicable in this situation. The electronic mineral conduction has already been discussed.

The presence of clay minerals lining pores may be important in another respect. The decrease in the conductivity of a bulk electrolyte after temperatures of about 350^{o}C has been ascribed to a decrease in the isobaric number density of charge carriers. There is a dynamic equilibrium between the clay surface cation areal density and the electrolyte cation volume density. Cations have the ability to desorb from the clay surface to restore the equilibrium disturbed by the progressive dilution of charge carriers by the decrease in their isobaric number density with increases in temperature. This mechanism is able to allay the reduction of electrolyte conductivity as temperatures increase. The extent to which this is possible depends upon the amount of adsorbed cations available for desorption which is dependent upon the efficiency of the surface in adsorbing cations and the surface area available for surface exchange. While all minerals adsorb cations, attracted to negative charges residing at their surfaces, some are much more efficient at it than others. Clays are particularly good. Clays lining pores also have much greater surface area to volume ratio than the other minerals in a rock due to their propensity to form complex lath and plate structures of fractal dimension between 2 and 3. As much as 99.9% of the internal surface area of a rock sample may be clays when the clay accounts for only 2% volume of the sample (British Petroleum data). There are more cations available for desorbtion in rocks with clay minerals lining the pores. The effective enhancement caused by desorbtion of cations in such rocks may be

an important secondary mechanism ensuring that more basic rocks, with clays lining their pores suffer less from the progressive decrease in electrolyte conduction at any given temperature than more acidic, clay free rocks.

Experiments on a graphite bearing garnet granulite showed some deviation from the behaviour of other basic granulite samples that were studied. The electrical conductivity of the graphite garnet granulite at lower crustal temperatures and pressures is consistent with values of lower crustal electrical conductivity obtained by MT methods (**Beamish, 1986**). The decrease in conductivity in the saturated rock with confining pressure was considerably less than expected. This is a result of an anomaly in the pressure variation of the dry rock. Measurements on the electrical conductivity of dry samples of the rock up to confining pressures of 0.4 GPa in the oil pressure vessel of the School of Environmental Sciences showed an increase in conductivity with pressure. This can be attributed to the progressive improvement of graphite conduction pathways as pressure acts upon the rock closing the pores. The experiments on rock samples taken perpendicular to the graphite foliation showed that this effect was sensitive to the distribution of graphite within the rock. Greater conductance and sensitivity of conductance to pressure occurred when the gross current flow was co-planar with the graphite foliations. It has been thought that the presence of large amounts of graphite bearing granulites were unlikely to be significant constituents of the lower crust. Recent work (**Frost et al., 1989; Peterson and Newton, 1989; Harris, 1989; Shankland, 1989**) is showing that such a model

cannot be excluded and may have significant advantages in describing the observed crustal conductivity values.

Graphite deposits in lower crustal rocks may be a lot more ubiquitious than has been thought **(Frost et al., 1989)**. They used advanced imaging techniques to identify very thin graphite films coating grain surfaces. They have calculated that such thin films, if present on grain surface over grains ranging in diameter from 0.1 to 10 mm, can produce lower crustal conductivities of the same magnitude as those measured in lower crustal rocks. They have noted that such high conductivities would not be observed in laboratory measurments due to the irreversible breaking of graphite films on decompressional straining of the rocks. They note that raising a high confining pressure would be unlikely to produce graphite reconnection and hence raise the electrical conductivity. This work has shown that reconnection can occur in graphite bearing lower crustal rocks of the type studied in this work.

The graphite conduction mechanism requires there to be some way of having graphite deposited in the lower crustal rocks and for the graphite to be able to produce high bulk rock conductivities. The mechanism for graphite film deposition in lower crustal rocks requires the presence of a carbon dioxide rich pore fluid, graphite may then be deposited by reduction or by oxyexsolution. Both of these mechanisms have wider implications upon the mineralogy of the bulk lower crust.

It is thought that the source of carbon dioxide is in the mantle **(Newton et al, 1980; Touret, 1971)**. Indeed fluid inclusions

found in many granulites are rich in carbon dioxide of an isotopic composition consistent with a mantle origin (Harris, 1989).

Fluxing of the lower crust with mantle derived carbon dioxide may have important mineralogical consequences for the rocks at depth. There is strong evidence that carbon dioxide rich fluids play an active role in granulite formation. Metamorphic equilibrium at the granulite facies boundary depends upon a number of factors including pressure, temperature and the nature of the pore saturating fluids. The displacement of aqueous pore fluids by carbon dioxide has been shown to be able to control granulite facies metamorphic boundaries in some areas (Janardhan et al., 1979). The presence of carbon dioxide promotes granulite formation and expels the water of dehydration subsequently produced. Such water is moved to higher layers due to gravity and the effect of carbon dioxide fluxing from the mantle. The overall effect would be to move the metamorphic boundary to a higher level where it is stable with the higher carbon dioxide flux (Jackson et al., 1988) and to produce upwardly mobile water. This process also enables the remaining granulites to be depleted in hydrophilic elements, a situation commonly seen in high grade metamorphic rocks. It may, therefore be said that carbon dioxide is most probably a factor in the generation of granulite facies rocks in the upper parts of the lower crust and a possible source for water which could saturate a layer above the zone of metamorphism. It should also be noted that there are other mechanisms by which granulite facies rocks can form such as by the metamorphism of already anhydrous lithologies (Lamb and

Valley, 1984) and the extraction from the parent rock of hygroscopic melts (Nesbitt, 1980). It is also interesting to note that the fluid carbon dioxide controlled metamorphic zones observed by Janardhan et al. (1979) and Jackson et al. (1988) were closely associated with shear zones: a point expanded upon in relation to the studies done relating to heating around such zones (Brodie and Rutter, 1987) later in this section.

Post-metamorphic carbon dioxide rich lower crustal rocks such as granulites may form graphite films on their surfaces on cooling by reduction of the carbon dioxide rich pore-fluid (Frost et al., 1989). Alternatively graphite may be formed by oxyexsolution, the mechanism by which titanomagnetite clears itself of titanium. Titanomagnetite and fluid carbon dioxide react to give magnetite, ilmenite and graphite (Fuhrman, 1988).

Granulites are depleted in those elements which preferentially partition into a coexisting melt and so it has often been said that magma has been extracted from such depleted assemblages during granulite facies metamorphism (Harris, 1989). Indeed water will readily partition into a melt, such that the final assemblage will be anhydrous as is required in granulite facies metamorphism; thus hygroscopic melt extraction may also be an important means by which granulite metamorphism occurs.

The partial melting of crustal rocks is often discounted due to the crustal temperature being too low except in the most active, young tectonic areas. Metamorphism by melt extraction has been limited to early in the evolution of a crust with most crustal differentiation occurring by this process. Such active areas as

the Cascade Mountains and the Basin and Range areas of North America have the expected high crustal conductivities. This has been attributed to reservoirs of hygroscopic melts and associated hot brines. Indeed the recent EMSLAB experiment has mapped the conductivity of the melt reservoirs underlying the Cascade Mountains (**Gough, 1989**).

Recent results from **Peterson and Newton (1989)** show that the temperature at which most common lower crustal rocks melt is reduced by the presence of carbon dioxide. Melt temperatures were depressed by as much as 50°C (to 760°)C at pressures of 1 GPa; pressures and temperatures believed to occur in high grade metamorphism in the lower parts of the crust (**Peterson and Newton, 1989**). Carbon dioxide is therefore consistent with the presence of deep level granulites depleted in magmaphilic elements as well as with the direct dehydration process in which rocks undergoing metamorphism are depleted in hydrophilic elements. The results of **Peterson and Newton (1989)** promote the possibility that partial melting can occur in certain parts of the hotter Phanerozoic lower crust providing that carbon dioxide is present. Such melts would enhance the electrical conductivity of the crust at deep levels. The presence of melts or layered intrusions of cooling, solidifying magma could explain some of the extensive seismic reflectivity of the lower crust. A partially molten crust would also be expected to exhibit the ductile, aseismic behaviour as is observed.

The graphite granulite samples (GGRAN1 to GGRAN4) studied in this work showed clearly that:

1. Reconnection occurred leading to high conductivities at raised confining pressures in the dry samples (GGRAN3 and GGRAN4).
2. The overall low confining pressure, low temperature saturated rock electrical conductivity was comparable with apparently graphite-free granulite saturated samples.
3. The reduction in the conductivity of the saturated graphite granulites is less than that for other metabasic rocks studied. This was attributed to the reconnection of graphite pathways in the saturated rocks.
4. The electrical conductivity of the graphite granulite samples increased more rapidly with temperature than the other metabasic rocks studied.
5. The raising of the electrical conductivity with temperature was smoother than that observed in the other metabasic rocks indicating the operation of fewer major conduction mechanisms over the whole temperature range. This is also well demonstrated by the single linear portion present in the inverse temperature variation graphs of the temperature variations of GGRAN1 and GGRAN2.
6. The physical anisotropy of the samples studied effected both the confining pressure and the temperature variations of the electrical conductivity of the dry and saturated samples. The samples cut parallel to the graphite foliations had larger temperature and confining pressure variations than those cut perpendicularly to the foliations. It is thought that

this is due to conduction being easier along graphite foliations due to greater graphite interconnectivity.

It may be said that several non pore fluid mechanisms may be important in the conductivity of rocks subjected to lower crustal conditions. These are;

1. Mineral conduction through the bulk of the rock especially through minerals containing a relatively large proportion of large cations e.g. magnetite, titanomagnetite, ferrous garnets and pyroxenes, and sodic and potassic hydrous hornblendes.
2. Mineral conduction through minerals of high conductivity such as graphite.
3. Conduction through and along the surfaces of alteration clay minerals lining pores and along grain boundaries.
4. Partial melting, if the lower crust is carbon dioxide saturated and the temperatures are sufficiently high.

The greater dependence of the electrical conductivity upon effective pressure in rocks of a more acidic composition was noted but was not as well developed as those cases observed by **Lee (1977)** and **Evans (1980)**. This may be due to the very small population of acidic rocks studied in this work or may be due to the maximum effective pressure of 0.2 GPa compared with that of 0.4 GPa used by **Lee (1977)** and **Evans (1980)**. The rocks of Lee and Evans were also slightly more quartz rich and while this factor is probably not sufficient to explain the effect on its own, all three of the above explanations could combine to explain

the dichotomy.

The variation of the electrical conductivity of the rocks with confining pressure at high temperatures shows a remarkably smaller variation than similar measurements taken at low temperatures. The results are inconsistent with information about the incompressibility of minerals with temperature (Handbook of Physical Constants, ed- Clark, S.P., 1966). One would expect the compressibility of a hard crystalline rock to mirror that of its constituent minerals. The compressibility of most major rock forming minerals increases with temperature by small amounts. Indeed the compressibility of a hard crystalline rock matrix made up of nominally incompressible materials might be expected to increase with temperature as a process of thermal expansion and external compression allows the grains to move relatively to each other into a more compact state, although such an effect would be likely to be extremely small for a low porosity rock. The explanation of the observed decrease in the conductivity/ confining pressure variation and apparent rock compressibility with increasing temperature must be sought in terms of more complex interactions in the bulk rock. Briefly, the electrical conductivity variation with confining pressure may be lessened due to:

1. Compensation by non-pore-fluid conduction mechanisms, which are less sensitive to changes in confining pressure, becoming more dominant at the higher temperatures.
2. Loss of conductivity of brines saturating the rock pores.
3. Thermal expansion of the pore-fluid keeping pores open

against the effect of confining pressures.

The results of this work have indicated that values of lower crustal conductivities have been underestimated. This has arisen because in the past conductivities have been found by extrapolating the variation of conductivity with increasing confining pressure at upper crustal conditions. This work has shown that the rate of decrease of conductivity with confining pressure is less at true lower crustal conditions leading to higher conductivities in the lower crust than were previously thought to exist. The results from the present study should give extrapolated values of electrical conductivity higher than those formerly deduced by around half an order of magnitude for lower crustal values. This is because the new data allow pressure extrapolation which takes into account the fact that the rate of change of electrical conductivity with confining pressure is itself a function of temperature.

The values of electrical conductivity at high pressure and temperature obtained in this study are in good agreement with the values of lower crustal conductivity obtained from MT measurements (Jones, 1981; 1982, Beamish, 1986). Saturated samples of the basic rocks that were studied (Table T9.6.; Figures F8.39. to F8.45., F8.47. and F8.49. to F8.50.) all had conductivities at lower crustal temperatures and pressures that were consistent with values obtained from MT studies. Neither of the more acidic rocks studied showed electrical conductivities at lower crustal conditions that would explain the observed MT values (Table T9.6.; Figures F8.46. and F8.48.).

The confining pressure variation of the electrical conductivity of saturated rocks at high temperature (Figures F8.63. to F8.69), the variation of conductivity with pressure and temperature for NaCl **(Quist and Marshall, 1968, 1969)** and the variation of conductivity with temperature and pore pressure for acidic rocks (Figures F8.46. and F8.48.), all suggest that electrolyte conduction in the rocks of the lower part of the lower crust is not controlling the values of the bulk electrical conductivity. Indeed these results indicate that the saturation state may be completely irrelevant to the values of electrical conductivity at these levels. Factors influencing the lack of electrolyte conduction control over the lower crustal conductivity include:

1. The permanent reduction in the porosity of rocks at high confining pressure (lower lower crustal levels) to the extent that all cracks and microcracks, and most larger HAR pores have been closed. The fractal nature of the pore surface ensures that porosity reduction is not complete even under extremely high pressures **(Gleick, 1988)** there will, however, be a limiting pore throat diameter below which the presence of fluid will not add to the conductance.
2. The rapid reduction in the electrical conductivity of the saturating electrolyte at temperatures above 350^{o}C, an effect more pronounced at low fluid pressures **(Quist and Marshall, 1968; 1969)**.
3. Desaturation by fluxing of the lower lower crustal rocks by mantle derived carbon dioxide.

The electrical conductivity/temperature variation for acidic rocks shows low conductivity at deep lower crustal temperatures due to the way that the electrical conductivity of the bulk pore fluid controls the saturated rock conductivity; the rock matrix being treatable as a non-conducting structure and the conductivity of the pore fluid becoming progressively lower at temperatures above around 300 to 400 ^{o}C. This suggests that electrolyte fluids, even if present in the pores of basic rocks in lower crust, would contribute little to the gross rock conductivity at these depths.

The observed high conductivities of the basic rock samples at lower crustal depths combined with the relatively low conductivities of the acidic rocks in these conditions thus indicate that the following major inferences may be made;

1. Electrolytic conduction is not the main mechanism of electrical conduction in the lower part of the lower continental crust.
2. The presence of water in the lower crust may be necessary, if present at all, only to activate clay mineral conduction.
3. Electrolytic conduction peaks at temperatures occurring in the upper part of the lower crust in most thermo-tectonic regions.
4. Mineral conduction increases steadily with temperature and is a much more dominant conduction mechanism in basic rocks than in the acidic rocks studied. High mineral conduction seems to be

correlated with composition, high mineral conduction occurring in basic rocks with:

- increased amounts of **spinels**, especially magnetite and titanomagnetite.
- increased amounts of **hydrated minerals** such as hornblendes.
- increased amounts of minerals containing **larger cations** such as ferrous garnets and pyroxenes, and sodic and potassic hornblendes.
- increased amounts of naturally **conductive minerals** such as graphite.

5. Rock conduction at depths correlated to the upper part of the lower crust may be due to the combined effects of:
   - the peak in the electrolyte conduction,
   - the increase in the mineral conduction (listed in point 5).
   - clay mineral surface conduction.
6. Rock conduction at depths correlated to the lower lower crust may be due to:
   - clay surface conduction activated by small amounts of moisture remaining in the lower crustal rocks or arising from H_2O-rich magmas in the upper mantle.
   - mineral conduction, especially graphite.

10.3. A MODEL FOR THE ELECTRICAL CONDUCTIVITY OF THE EARTH'S CONTINENTAL CRUST

The broad features of the crust will be examined before any explanation will be attempted. The crust can be broadly categorised into the upper crust and the lower crust. The upper crust is the better understood of the two due to greater opportunities for sampling its physical and chemical properties. The upper crust is electrically resistive, although containing high conductivity zones in some areas. It is relatively seismically transparent, contains virtually all the intra-continental earthquake centres and responds brittly to anisotropic stresses. The lower crust, in comparison, is more electrically conductive, seismically reflective, aseismic, responds to stress in a more ductile manner and has higher seismic velocities (Gough, 1986). The lower crust is apparently composed of an upper part of high electrical conductivity and a lower layer of moderately raised conductivity, although ambiguity in the magneto-telluric inversion techniques makes it possible that a very thick high conductivity layer is present or a very high conductivity upper layer of greater thickness is masking a relatively low conductivity lower zone (Hyndman and Shearer, 1989). Despite uncertainties in the magneto-telluric interpretation there is no doubt that the lower crust is more conductive than would be expected from a dry lower crust of anhydrous mineralogy with a low porosity induced by high confining pressures. The higher conductivity of the lower continental crust seems to be of global extent although there is also a correlation to geological age. The conductivity of the upper crust is, for comparison, 10^{-4} - 10^{-5} S m^{-1}. This is too

high to be due to dry rocks and is explainable by conduction through electrolytes saturating the rock pores. The highest conductivities of the lower crust (10^{-1} - 10^{-2} S m^{-1}) occur in its upper parts in Phanerozoic mobile belts and presently active tectonic areas. These areas have lower lower crustal conductivities around 10^{-3} S m^{-1}. There are mechanisms that can explain this conductivity distribution and these will be detailed later in this section. What is more surprising, perhaps, is the more moderate rises in the electrical conductivity of the whole of the lower crust of Pre-Cambrian stable shield areas (10^{-3} S m^{-1}) **(Shankland and Ander, 1983; Adam, 1976; 1987; Haak and Hutton, 1986; Jones, 1981; Hjelt, 1988; Gough, 1989)**. The layers of high electrical conductivity in the upper part of the lower crust can be correlated to the onset of the seismically reflective lower crust **(Gough, 1986; Jones, 1987; Hyndman and Shearer, 1989)**

There are several mechanisms which may have a part to play in the electrical conductivity of the continental crust:

- Electrolyte fluids **(Hyndman and Shearer, 1989; Bailey et al., 1989)**
- Graphite **(Keller and Frischknecht, 1986; Frost et al., 1989; Shankland, 1989)**
- Metallic oxides **(Lebedev and Shepel, 1981; Fuhrman et al., 1988)**
- Sulphur **(Goldsmith, 1976)**
- Carbon dioxide **(Wendlandt, 1981; Harris, 1989; Jackson et al., 1988; Peterson and Newton, 1989;**

Frost et al., 1989)

- Hydrated minerals

  serpentinites (Parkhomenko, 1982)

  clays (Jackson, 1984)

- Partial melting (Jackson et al., 1988)
- Shear zones (Brodie and Rutter, 1987; McCaig, 1989; Harris, 1989).

All these have been discussed in detail in the previous chapters. Any model of the electrical conductivity of the continental crust needs to bear in mind all the possible mechanisms and all the observations aforementioned.

Figure F10.1. shows the broad features of the proposed crustal model. It relies on the conductivity being provided at different depths by admixtures of the following mechanisms:

1. Electrolytic conduction - increasing to a maximum in the mid-crust and subsequently decreasing to negligable levels.
2. Mineral conduction - increasing with depth and temperature, but increasing at a much greater rate in basic rocks compared with acidic rocks.
3. The presence of graphite in the lower lower crustal rocks.
4. Possible partial melting in the lowermost parts of the lower crust in young crust and also in older crust in the presence of carbon dioxide.
5. Fluxing of the lower contiental crust with carbon dioxide from the mantle.

The model may be split into layers but will be described as a

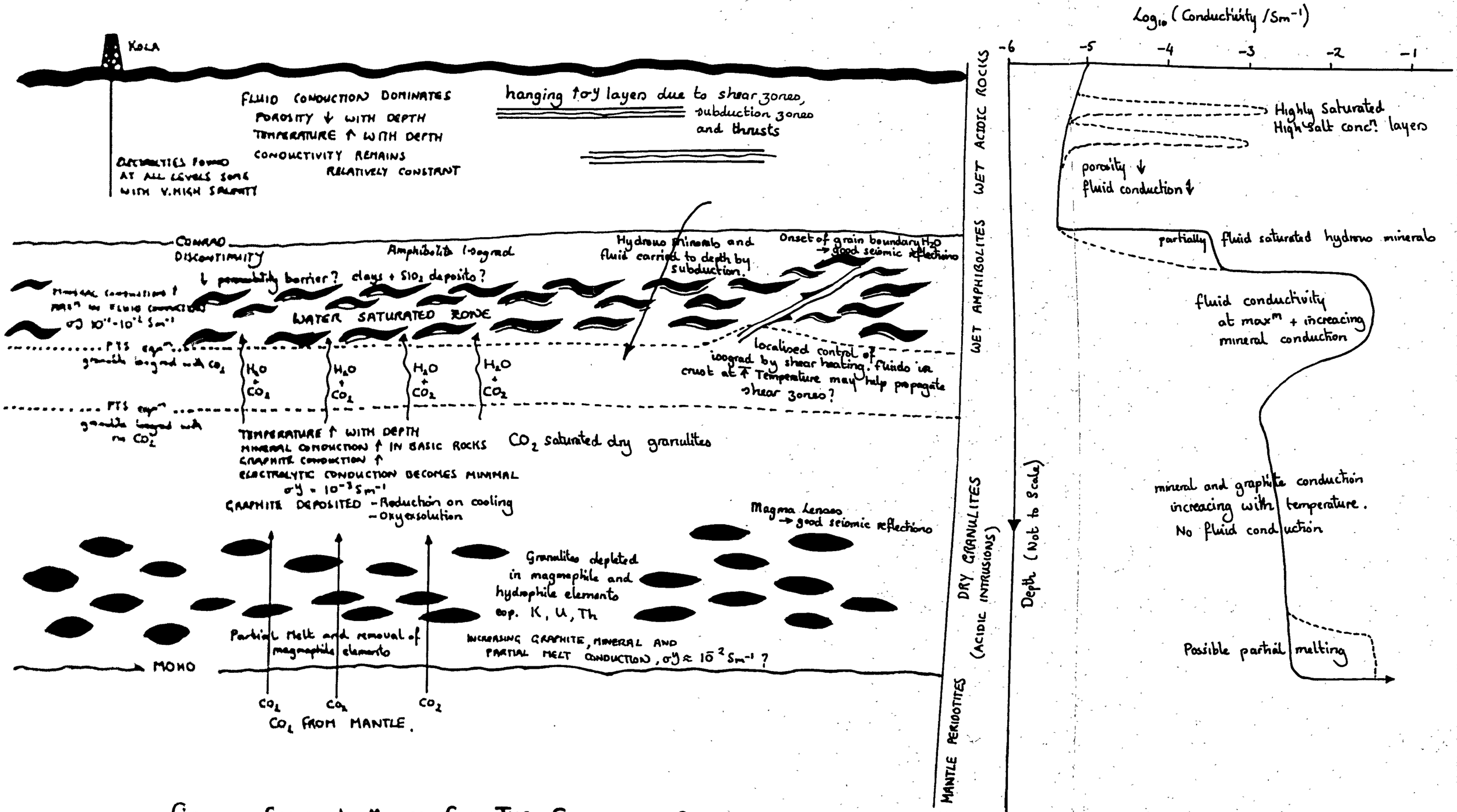


FIGURE F10.1 A MODEL FOR THE ELECTRICAL CONDUCTIVITY OF THE CONTINENTAL CRUST.

whole here. Likewise it is easier to pigeonhole the evolution of a crust into sampled times, whereas the true case is a continuous process. Hopefully the sense of such a process is retained in the following description.

The topmost layer of the crust has the greatest porosity and a significant percent of the porosity in the form of low aspect ratio cracks. It is saturated with water of approximately the same concentration as seawater. Water circulation within this part of the crust is easier and swifter than in any of the other layers. The high conductivity is probably due to high fracture porosity (Brace, 1971; Dvorak, 1975) and leads to high bulk saturated rock conductivities. The typical conductivities of this layer are wideranging (10^{-5} - 10^{-2} S m^{-1}) indicating its inhomogenous nature. It occupies only the first few kilometers of crust. About the next 20 km of crust can be treated as the first major unit (the upper crust). It has typical conductivities ranging between 10^{-6} - 10^{-5} S m^{-1} and is a zone of unweathered fluid saturated rock predominated by acidic rocks with considerable lateral and vertical inhomogeneities. Porosity is never as great as in the topmost layer and progressively decreases as depth is increased (Figures F8.1. to 8.12). Closure of small aspect ratio cracks at shallow depths progresses to include larger aspect ratio pores as depth increases. The decrease in rock conductivity from this process is compensated for by the increase in conductivity due to increasing temperature as depth is increased (Figures F8.39 to F8.50.). Nevertheless conductivity is never as high as in the surface layer. There may be thin layers of higher conductivity (up to 10^{-4} S m^{-1}) within

this layer (Bailey et al., 1989). Such areas may indicate local high brine saturation, high brine concentrations or greater dilatancy and may be correlatable to shear zones, subduction zones and thrusts (McCaig, 1989; Hyndman and Shearer, 1989).

The base of the upper crust is usually correlatable to a visible, seismically defined, Conrad discontinuity in the seismic velocities. When this is not present a figure for seismic velocity greater than 6.5 km s^{-1} is taken as indicating lower crustal rocks. The transition has been taken to be the change from acidic dominated mineralogies in the upper crust to hydrated basic dominated ones in the upper part of the lower crust (usually amphibolite/hornblendeite).

As the depth into an amphibolite facies rock increases, the electrical conductivity of the rock would be expected to increase in line with the increase in the bulk fluid conductivity with temperature minus that decrease due to the progressive closure of porosities by confining pressure increases. In addition, the conductivity would also rise due to the steady increase in mineral conduction with temperature, which, by these depths, is beginning to have a significant effect on the overall conductivity of the rock because;

1. the temperature is increasing
2. the amphibolite is a hydrous rock facies
3. amphibolites commonly contain increased proportions of ferrous pyroxenes, garnets and magnetite-like minerals that have been seen to give the metabasic rocks a higher conductivity that acidic rocks.

In a crust there may be such a level where the pressure, temperature and saturation are such that granulite facies metamorphism can take place if a suitable lower grade facies is present. These reactions can take place by several mechanisms such as direct dehydration, melt extraction, and metamorphism of already anhydrous facies rocks. Whether these reactions take place depends entirely upon the values of pressure, temperature, saturation and the presence of a 'reactant'. A level may exist where these factors are all such that equilibrium occurs and the rate of the reaction is product-controlled. the production of water from such reactions can alay further reaction. If it is not removed from the vicinity, the water will quickly recombine to form hydrous minerals (Yardley, 1981; Wood and Walther, 1983).

Large amounts of carbon dioxide have been found in inclusions in samples of rock from the lower crust (Jackson et al., 1988; Janardhan et al., 1979; Newton et al., 1980). The carbon dioxide appears to have come fom the mantle as has already been discussed. Fluxing of the lower crust by mantle derived carbon dioxide can upset the metamorphic equilibrium by displacing water and hence driving the reaction to produce more water and to begin at lower temperatures. The greater production of water is removed by the carbon dioxide. Thus fluxing by carbon dioxide promotes granulite facies metamorphism and allows the level at which such reactions can begin to migrate to higher levels in the crust. The removal of water of dehydration by the carbon dioxide ensures that the dehydration reactions go to completion. The removal of hydrophilic elements, especially those which produce radiogenic heat, occurs in association with the removal of water.

The water is displaced to levels closer to the surface than those at which granulite facies reaction begin to occur. This is ensured by the feedback mechanism of too great a water saturation arresting further dehydration reactions. This is a mechanism for the production of water saturating hydrous amphibolites at temperatures between 300 to 500 ^{o}C if the water can be kept at these levels.

The mechanism of metamorphism causes release of water and may lead to significant solution of conducting ions in the conductive fluid enhancing the conductivity of the electrolyte. The large scale and incipient nature of the metamorphism ensures that the rock is completely saturated, along grain surfaces with a low dihedral angle **(Hyndman and Shearer, 1989)** with fluid which generates a pore pressure acting in direct opposition to the confining pressure opening pores and increasing the connectivity of the conducting fluid (Figures F8.15 to F8.26.). The mechanism of metamorphism may also alter the porosity of the rock. Fluid generated at this level increases the conductivity, particularly as the fluid has its maximum conductivity at these temperatures (Figure F8.46. and F8.48.).

The fluids formed in the dehydration reactions will have a natural tendency to rise in the crust due to the differential gravitational forces acting upon the fluids and the rock matrix because of their different densities. This effect combined with the dewatering effect of the carbon dioxide causes the fluids to percolate upwards and be lost to the upper crust if no barrier is available to stop them. The rate that this occurs is dependant

upon the permeability of the rocks. Calculations by **Hyndman and Shearer (1989)** indicate that permeabilities at the extreme of those measured in laboratory studies on hard, low porosity, crystalline metamorphic and igneous rocks **(Trimmer et al., 1980)** are necessary to keep compaction released water confined to the upper part of the lower crust long enough to enable a highly conducting layer to form. These permeabilities criteria can, however, be relaxed if there is a source of water present on time scales less than the 100 Ma assumed by Hyndman and Shearer for compaction generated water. The metamorphically generated water may provide such a lower timescale source of water. Produced water may also be retained in the lower crust if there is a zone of particularly low permeability above it. Such a capping layer has been proposed **(Etheridge et al., 1983)** to occur due to precipitation of silicate materials, especially clays, from a hot saturated aqueous pore-fluid of metamorphic genesis. The capping layer is also cited by **Jones (1987)** to trap water in an electrically conducting lower crustal water saturated layer in his electrical model for the continental crust. It must be said that the capping layer is entirely speculative and may not even be needed if the metamorphic water source is sufficiently productive and the rocks of the upper part of the lower crust have similar permeabilities to those measured in the laboratory **(Trimmer et al., 1980)**.

The above discussion has put the case for a mid-crustal amphibolite layer at least partially saturated over a proportion of its thickness. The temperature at such depths coincides with the maxima in the bulk electrical conductivity of aqueous pore brines. It is therefore proposed that a significant factor in

the mid-crustal high electrical conductivity layers is the presence of hot brines of metamorphic genesis promoted by carbon dioxide fluxing. This, however, is far from the complete picture of the processes acting at depths to influence the crustal conductivity and mineralogy. The continued rise in temperature with depth, an increase more pronounced in younger, more tectonically active areas, causes mineral conduction to begin to become significant at mid-crustal levels. Solid-state mineral conduction occurs in the hydrous, large cation mineralogy of amphibolite facies rocks. If clays are present then these may add to the conductivity by conduction through their matrix and also by surface conduction. Graphite may also be present in a amounts sufficient to significantly add to the bulk electrical conductivity of the rock at mid-crustal depths. It is therefore envisaged that the topmost part of the high electrical conductivity layer is explainable in terms of a mixture of conduction mechanisms including the maxima in the pore-fluid conduction and the major addition of mineral conduction increased by the mid-crustal high temperatures. The conductivity of the upper part of the lower crust in Phanerozoic areas ranges from 10^{-1} - 10^{-2} S m^{-1} and in older more stable areas from 10^{-2} - 10^{-3} S m^{-1}. The extent of the layer seems to be about 20 - 30 km depth in the Phanerozoic areas whereas the whole of the lower crust may be involved in stable areas and the mechanisms for conduction may be significantly changed.

The presence of a layer of grain boundary water saturated amphibolite facies rocks at mid-crustal levels would be expected to show up as a strong seismic reflector especially if

interspersed with local areas high in carbon dioxide and mainly dewatered. The Conrad discontinuity existing at these depths at many places around the world may be correlated to such a water saturated layer.

Completion of granulite facies metamorphism leaves a hard, compact, carbon dioxide saturated, anhydrous rock. Porosity has been reduced to a very small value by the increasing pressure. The majority of fluids exist along the grain boundary porosities. An extremely low conductivity would be expected for this layer if conduction were due to electrolytic fluid conduction alone due to the low porosity and low electrolyte electrical conductivity at lower lower crustal temperatures. The temperature of the lower part of the lower crust ensures that the conductivity of intrinsically more conductive minerals such as magnetite, titanomagnetite or ilmenite is significantly large compared with their conduction at lower temperatures. Conduction can occur via various types of temperature-activated mineral conduction such as;

- spinels, especially magnetite and titanomagnetite,
- minerals containing a greater proportion of larger cations (Parkhomenkho, 1982),
- naturally highly conductive minerals such as graphite.

This in itself is not be sufficient to account for the conductivities reported from MT measurements (Jones, 1987).

This part of the lower crust may be rich in carbon dioxide as has already been said. Such a situation may give rise to graphite

deposition by reduction on cooling or by oxyexsolution (**Frost et al., 1989; Fuhrman et al., 1988**). The presence of graphite can account for high bulk rock electrical conductivities even if at small weight percentage of graphite if the graphite is distributed as thin films on grain surfaces. Graphite deposition as a result of reduction of carbon dioxide would be in the form of thin grain coating films. Oxyexsolution would, however, tend to produce graphite in the spinel-rich areas of the rock leading to a more granular graphite distribution similar to that found in the rock samples studied in this work (GGRAN1 - GGRAN4). The process of oxyexsolution would produce graphite that would be expected to be associated with the presence of ilmenite and magnetite, both minerals with high solid state mineral conductivity at lower lower crustal temperatures. The electrical conductivity of the lower part of the lower crust is proposed to be due to mineral conduction at high temperatures in which graphite takes a leading role.

It has been previously noted that there is a moderate increase in the electrical conductivity at mid-crustal levels in Pre-cambrian cratons and associated stable areas compared with the dramatic high conductivity layers observed in Phanerozoic mobile belts and presently active tectonic areas (**Hyndman and Shearer, 1989**). This behaviour is not explained by models such as that of **Hyndman and Hyndman (1968)**, and the model of **Jones (1987)** relies on non-graphite, inherent mineral conduction to provide moderately raised conductivities at high temperatures at lower lower crustal depths. This assumption may not necessarily hold true although the samples studied in this work, both graphite-rich and graphite free basic granulites, had sufficiently high conductivities at

lower crustal temperatures to explain the observed field conductivities.

Hyndman and Hyndman (1968) postulated that areas of young thickened crust arising from deformation of water saturated geosynclinal sediments would be heated by a combination of ex-mantle heat-flow and intra-crustal radioactive heating from elements found in naturally inceased abundances in the surficial sediments. Such heating caused partial melting in the water saturated lower levels of the thickened crustal slab, the granitic melt of which would be mobile and move towards the surface to give a granitic upper layer. There would be granulite facies metamorphism and hence dehydration of the remaining lower level rocks, and leaving a restite preferentially enriched with more basic rocks and depleted of the radioactive elements that, being soluble in water and melt, move with the granitic melt. The mobility of the heat producing elements ensures that the whole crust is progressively subjected to such a differentiation 'by-fire' from the base upwards. The electrical outcome of such a model is a high electrical conductivity in the young undifferentiated greenschist to amphibolite facies crust which, over time becomes less thick due to the increasing transformation of the lower parts of the crustal slab to anhydrous low conducting granulite rocks. The dehydration horizon moves upwards with the migration of the heat producing elements dissolved in the dehydration-released water. Ultimately the process ends with all of the crust differentated into a granitic and high grade metamorphic acidic rocks high in radioactive elements lying above an extensive anhydrous low conducting

granulite facies layer. The stable shield crust so produced has generally low conductivities at all depths and shows no high conductivity layer due to there being no saturated hydrous lower grade layer of rock remaining.

The moderate increase in the conductivity of the lower crust in stable shield areas is not explained by the Hyndman and Hyndman model. The graphite rich crust scenario described in this thesis, can result in a stable shield having moderately raised electrical conductivities due to non-graphite mineral conduction and graphite conduction. As the young crust becomes progressively older, hydrophilic, radiogenic heat-producing elements are removed from the lower crust and transported to the upper crust by water. The crust as a whole cools down and graphite precipitates by the reduction of carbon dioxide. The very old stable crust has a lower crust that has been completely transformed to anhydrous graphite bearing granulites. Most of the water from metamorphism or any other source has been lost to the upper crust. All the water that remains may be the water vapour of mantle origin coexisting with the mantle derived carbon dioxide in small grain boundary porosities. Ultimately the crust consists almost entirely of a basic, relatively dense, graphite bearing dry granulite lower crust and a lighter, water saturated, acidic upper crust preferentially enriched in heat producing minerals. A thin amphibolite band separating the two layers may also continue to exist. The old stable lower crust then has a moderately raised electrical conductivity due to the presence of high temperature mineral conduction especially in the form of graphite.

The model by Gough (1986) is based upon the analysis of stress regimes in the upper and lower crust. It well describes the relative magnitudes of conductivity in the crust but does so in a qualitative manner. Unfortunately it is only completely applicable in areas where the upper crust is undergoing lateral compression. He points out, however, that fluid saturation of parts of the lower crust may be due to the subduction of water saturated sediments (Fyfe, 1981; Fyfe, 1986) and that such water, if distributed intergranularly, would account for the observed ductile behaviour of lower crustal rocks. He points out that water at lower crustal levels, present from whatever source, would be forced to occupy grain boundary porosities by the high and almost homogeneous high lithostatic pressures in just this way. The cooler upper crust would be exposed to relatively greater differences in least principal stresses and would have water more confined to macroscopic pores due to closure of pore connections due to the larger of the principal stresses. This effect would be most pronounced in rocks subjected to highly differential compressive stress fields similar to those likely to be encountered in the upper crustal tectonically active belts. Such a system would be more prone to elastic strain/ brittle failure which agrees well with the observed distribution of intracontinental earthquake epicentres (Chen and Molnar, 1983).

The perfusion of high-grade shear zones in the deep crust with water may be an important source of high conductivity although their effect may be more localised than the generalisations that have been described above. Work done by Brodie and Rutter (1987) towards the mapping and description of high-strain high-

temperature shear zones in the Ivrea Zone showed that the tectonics of the lower crust did not have to be viewed entirely as a result of an all pervasive ductile movement. They described localised high-temperature shear zones and how these gave rise to extension of the Ivrea Zone amphibolite and granulite ortho- and para-gneisses. The zones could be seen as well localised zones of shear strain and can be treated as discontinuities in the rock, in a similar way to faults, because their widths were small compared with their visible extent and inferred continuity. Extensional faulting, similar to that present in the Ivrea Zone, progresses in a few large discrete faults. Each movement juxtaposes parts of the hanging wall against higher temperature foot-wall of large extent and hence great thermal capacity. The heating of the hanging-wall by the foot-wall causes a temperature step to develop in the vicinity of the shear zone. When combined with the the shear heating associated with the strain developing the overall temperature increase has been calculated by **Brodie and Rutter (1987)** to be in excess of 50^{o}C, sufficient to induce a change from amphibolite to granulite facies with the resulting liberation of water which would then pervade the shear zone. The shear zone would delimit the facies transition and the facies transition would mark the position of rock fully pervaded with a continuous film of highly conductive fluid which could lead to an abrupt increase in electrical conductivity at this point. The effect is dependent upon the discontinuous temperature variation that is produced by the shear zone in the region of hydrated facies and initial temperatures just below the zone being such that the undisturbed amphibolite/granulite isograd occurs there. In short the shear zone must cut or closely approach the dehydration isograd and the temperature step be enough to cause

the metamorphism of hydrated amphibolite to dehydrated granulite with the resultant migration of the isograd up towards the shear zone and the release of water. It can be seen that this is a mechanism by which a water saturated, highly conducting horizon may be 'forced' by the presence of a shear zone. It is also tempting to speculate that the propagation of extensional faulting by high temperature shear zones might be greatly eased and focussed by the heating occurring in the zone due to the two mechanism described by **Brodie and Rutter (1987)** and also by the lubricating effect of the water released by such temperature steps occurring near the dehydration isograd. The combination of these effects would lead to a subsequent preference for shear along the isograd. **Brodie and Rutter (1987)** have noted the presence of a high temperature shear zone cutting this particular isograd in the Valle d'Ossola area of the Ivrea Zone.

Brief mention should also be made of water taken to depth in subducting plates and water saturated sediments, and along shear zones and thrusts **(McCaig, 1989; Kurtz, DeLaurier and Gupta, 1986; Clowes et al., 1987;, Green et al., 1987; Waff, 1988)**. There are conducting zones associated with such phenomena and these highly conducting dipping layers are correlated to dipping seismic reflectors. It seems that the presence of fluid saturated zones can produce good seismic reflectors in these areas.

The argument above has concentrated upon explaining the conductivity in a pre-differentiated crust. In reality differentiation occurs contemporaneously with the situation

described in the last fifteen pages. While some mass transport and differentiation can occur by hydrophilic transport, the huge volumes of material needed to differentiate the crust into the broadly acidic and basic final product would almost certainly require partial melting and magma transport to be a major effect at least in the hotter early evolution of the crust.

Crustal differentiation is generally accepted to occur by percolation of hot acidic melts from the lower parts of the crust where the mantle heat flux and the initial presence of radiogenic heating from isotopes of potassium, uranium and thorium have resulted in a sufficiently high temperature (**Taylor and McLennon, 1984**).. The process is self-limiting due to the gradual loss of the lighter, lower melting point fractions of the undifferentiated rock together with the radiogenic heat producing elements and water which preferentially dissolve in the melts. The lower crust becomes progressively cooler and of basic composition while the radiogenic element-rich melts solidify in the upper crust. The lower crust becomes depleted in the radioactive elements and water. The presence of water in the initial assemblage has been shown to reduce the temperature at which the undifferentiated rocks melt. Another fluid capable of doing this is carbon dioxide (**Peterson and Newton, 1989**). Indeed the recent results of these workers show that the melting point of crustal rocks can be reduced by as much as 50^{o}C enabling the possibility that partial melting could not only occur in relatively young and active areas but also in the deeper parts of relatively old crust. The scenario involves the partial melting and differentiation of the crust early in its evolution but also later on near the Moho. Such areas of deep, late partial melting

could solidify at slightly higher and cooler levels of the older lower crust to produce layered intrusions and lenses of material capable of producing strong seismic reflections of the type observed in the lower crust **(Matthews and Cheadle, 1986)**.

In brief, the model proposed in this work can explain;

1. Upper crustal conduction by pore fluid conduction in acidic rocks - decreasing with confining pressure as porosity decreases and increasing with temperature.
2. Upper lower crustal high conductivity in Phanerozoic and young active areas by;
   - pore fluid electrolyte electrical conduction being at a maximum,
   - water resulting from metamorphic, carbon dioxide promoted reactions and subducted waters in various ways,
   - mineral conduction at raised temperatures in;
     - hydrous minerals with large cations
     - magnetite-like minerals
     - clays
     - perhaps graphite,
   - partial melting in very young or particularly active areas or in the deepest parts of Phanerozoic crusts if temperatures at these levels are sufficient.
3. Lower lower crustal moderately raised electrical conductivities in Phanerozoic and younger areas due to mineral conduction at higher temperatures than in point 2. above with the significant addition of graphite

conduction but the absence of pore fluid conduction.

4. Total lower crustal moderately raised conductivity in Pre-cambrian and associated old stable areas by evolution of the Phanerzoic lower lower crust to encompass the whole lower crust with the loss of the water saturated Phanerozoic upper lower crust.

5. The concentration of intra-continental epicentres in the upper crust due to the upper crust being pore fluid saturated, cold and under differential stress regimes. The lower crust, by comparison, is at least partially saturated with grain boundary water and high pressure carbon dioxide, hot, more ductile and subject to a more homogenous stress field.

6. The concentration of seismic reflectors in the upper lower crust due to water saturated amphibolite zones alternating with areas of greater carbon dioxide saturation. Seismic reflectors in the deeper lower crust may be due to layered acidic intrusions in relatively denser basic restite. Such intrusions could be formed from partially molten basal crust with melting promoted by CO_2.

7. The seismic transparency of the upper crust due to the lack of density differentiation between allied blocks and layers of magma derived acidic upper crust on the scale of resolution of deep crust seismic studies.

The model has broad similarities with the models of **Gough (1986)**, **Jones (1987)** and **Hyndman and Hyndman (1968)** but has the advantage of being able to explain the moderate rises in the conductivity of the lower lower crust in Phanerozoic and younger areas and the

moderate rises in the conductivity of the stable Pre-cambrian total lower crust which these models cannot.

The set of rocks studied in this work is far from exhaustive. More samples need to be studied in various saturation states at high temperatures before more specific conclusions can be drawn as to the true electrical and compositional state of the lower crust. It may prove that many sufficiently basic saturated rocks have the correct electrical properties to be considered as possible major components of the lower crust. Whatever the final outcome, this work has shown that electrical measurements are possible at lower crustal temperatures on completely saturated rocks and that these measurements can help clarify the electrical properties of the lower continental crust.

PUBLICATIONS

PAPER 1: GLOVER, P.W.J. et al., 1988
The Measurement of Electrical Resistivity of Rocks under Simulated Crustal Conditions.
In High Pressure Geosciences and Materials Synthesis, Ed: Vollstadt, H., pp 46-49, Academie-Verlag, Berlin.

PAPER 2: GLOVER, P.W.J. et al., 1989
The Measurement of Saturated Rock Electrical Conductivity at Lower Crustal Temperatures and High Pressures.
Given at the 12th AIRAPT Conference, 1989,
Submitted to the Journal of High Pressure Physics.

PAPER 3: GLOVER, P.W.J. et al., 1989
Measurements of Saturated Basic and Acidic Rock Electrical Conductivities at Lower Crustal Temperatures and High Pressures.
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THE MEASUREMENT OF ELECTRICAL RESISTIVITY OF ROCKS UNDER SIMULATED CRUSTAL CONDITIONS

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The observed field determinations of crustal conductivity depth profiles show anomalously high conductivities in the lower crust. This has yet to be explained by a consistent theory and interpretation of field results is difficult due to the lack of relevant laboratory-based conductivity measurements. Although there are laboratory measurements of rock conductivity of the upper crustal regime[1] no laboratory measurements on saturated rocks are available at lower crustal temperatures and pressures. Measurements of saturated rock cores at temperatures of up to 800 ^{o}C and pressures of up to 1 GPa while enabling the variation of pore pressure independently of confining pressure poses considerable problems in designing an experimental cell. An initial design of experimental cell is presented. Its design will be discussed with special attention given to problems obtaining an insulated two terminal conductivity measuring electrode arrangement at high temperatures while retaining a jacketed sample by the use of a guard-ring electrode arrangement. The stress in the cell and design of measurement seals will also be discussed. It is hoped that preliminary results from measurements on amphibolite and granulite facies rocks will also be presented.

[1]Lee, C.D., Vine, F.J. and Ross, R.G., Geophysics J.R. astr. Soc. (1983), 72, 353-371.

THE MEASUREMENT OF ELECTRICAL RESISTIVITY OF ROCKS UNDER SIMULATED CRUSTAL CONDITIONS

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ABSTRACT

The observed field determinations of crustal conductivity depth profiles show anomalously high conductivities in the lower crust. This has yet to be explained by a consistent theory and interpretation of field results is difficult due to the lack of relevant laboratory-based conductivity measurements. Although there are laboratory measurements of rock conductivity of the upper crustal regime[4] no laboratory measurements on saturated rocks are available at lower crustal temperatures and pressures. Measurements of saturated rock cores at temperatures of up to 800°C and pressures up to 1 GPa while enabling the variation of pore pressure independently of confining pressure poses considerable problems in designing an experimental cell. An initial design of experimental cell is presented. Its design will be discussed with special attention to problems obtaining an insulated two terminal conductivity measuring electrode arrangement at high temperatures while retaining a jacketed sample by the use of a guard-ring electrode arrangement. The design of seals will also be discussed.

INTRODUCTION

Field studies commonly reveal a zone of high electrical conductivity within the Earth's lower crust. Various attempts have been made at explaining such behaviour but all are hampered by the lack of knowledge of the electrical behaviour of rock types found at such depths under high pressures and temperatures. Direct measurements of the electrical properties of rocks in situ is not possible so laboratory experiments under artificially created pressure temperature conditions are of great value. One of the possible explanations of the lower crustal high conductivity layer involves the presence of an aqueous electrolyte. It is therefore important that such measurements are possible on saturated rocks with the ability to vary the pore pressure independently of the overburden pressure. Up until now such work has been carried out up to the upper crustal conditions of 0.4 GPa and 250°C[1,2,3,4]. The value of temperature and pressure in the lower crust is higher than this in most areas. The range of temperatures and pressures needs to be extended to 800°C and 1 GPa respectively. This is the primary aim of this work and demands the design of a measurement cell that will withstand such an extreme environment.

EXPERIMENTAL CELL DESIGN

An initial measurement cell has been made (Fig. 1). Its design has been dictated by the high pressures and especially temperatures at which it is planned to be used. The ability to vary the pore pressure of a saturated rock sample independently of the

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confining pressure requires that the rock is surrounded by a sleeve that will separate the external pressure transmitting medium from the internally pressurised electrolyte solution. This sleeve must be sufficiently deformable to transmit the confining pressure to the rock sample. For temperatures up to around 250°C a PTFE sleeve has been used[3]. The electrical insulating properties of the PTFE enables the end pieces of the cell to act as electrodes. Extending the range of temperatures at which such a conductivity cell can operate entails the replacement of the PTFE sleeve with another deformable material that can withstand high temperatures. Unfortunately there is no such material that is an electrical insulator as well as fulfilling these requirements. A copper cylinder, 250 μm thick, has been chosen to sleeve the sample. The use of an electrically conducting copper sleeve raises two major problems that need to be solved independently:

a) Electrodes placed at each end of a cylindrical sample need to be set in a suitable electrical insulator that can be sealed to the sleeve and electrode leadthroughs.
b) Leakage currents shunted around the sample by the copper sleeve give rise to a measurement of apparent conductivity that is considerably greater than the true value of the rock conductivity.

The next two sections deal with these two problems in this order.

MATERIALS AND SEALING CONSIDERATIONS

The use of a copper sleeve has dictated many of the requirements for other parts of the cell. Reference to Figure 1 shows the need for setting the electrodes in a block insulator at each end of the cell. Such electrode holders are required to have high electrical resistance, zero connective porosity, high compressive strength, be refractory to 850°C and have an ability to be sealed to the copper sleeve and electrode leadthroughs. Pure recrystalised alumina manufactured by isostatic pressing by Degussa fulfils these requirements and has been tested to 0.4 GPa. A computer model has indicated that no failure should occur due to compressive or tangential stresses at confining pressures up to 1 GPa. Tests on helium leak detector indicate zero porosity for alumina made by isostatic pressing but not for alumina that is slip-cast. The electrode leadthroughs need to be sealed to the alumina to isolate the confining pressure medium from the pore pressure transmitting aqueous NaCl solution. Niobium was chosen as a suitable material for the leadthroughs as it has almost exactly the same linear expansivity as the alumina. This enables the use of a simple cone seal. The initial seal is made by diamond lapping the niobium into a self-locking taper that has been diamond ground into the alumina. Since the confining pressure is always greater than or equal to the pore pressure the seal remains tight at high pressures. The similar and small linear expansivities of both alumina and the niobium ensures the seal remains tight at high temperatures. Copper slip is used as a gasketting material to improve sealing. Electrodes are made from stainless steel to withstand attack by the pore pressure solution, although they may have to be fabricated from another material to withstand attack by hot electrolytes. The electrodes are machined slightly too small to accommodate relative thermal expansion with respect to the alumina. They are held against the sample by a light spring, which, with the presence of a surface layer of electrolyte on the sample, in the case of saturated samples, provides good electrical connection. The alumina/copper sleeve seal shown in Figure 1 is a modification of the

design of H. Jolly, School of Mathematics and Physics, University of East Anglia. The seal is initially made by swaging the copper sleeve into a groove in the alumina using a rotating wheel tool. A circlet of copper wire is laid in the copper groove and held in place by four jaws. The four stainless steel jaws are tightened together by sliding a tapered collar over them by the use of an assembly tool. The seal will improve at high temperatures if the collar is made from a material of less linear expansivity than the stainless steel and copper because the expansion of the copper sleeve/wire and steel jaws is constrained between the less expansive collar and alumina. Leak tests on this seal show no helium leakage when tested on a helium leak detector and no penetration of high pressure transmitting oil when held at 0.4 GPa for 24 hours in a hydrostatic oil pressure vessel.

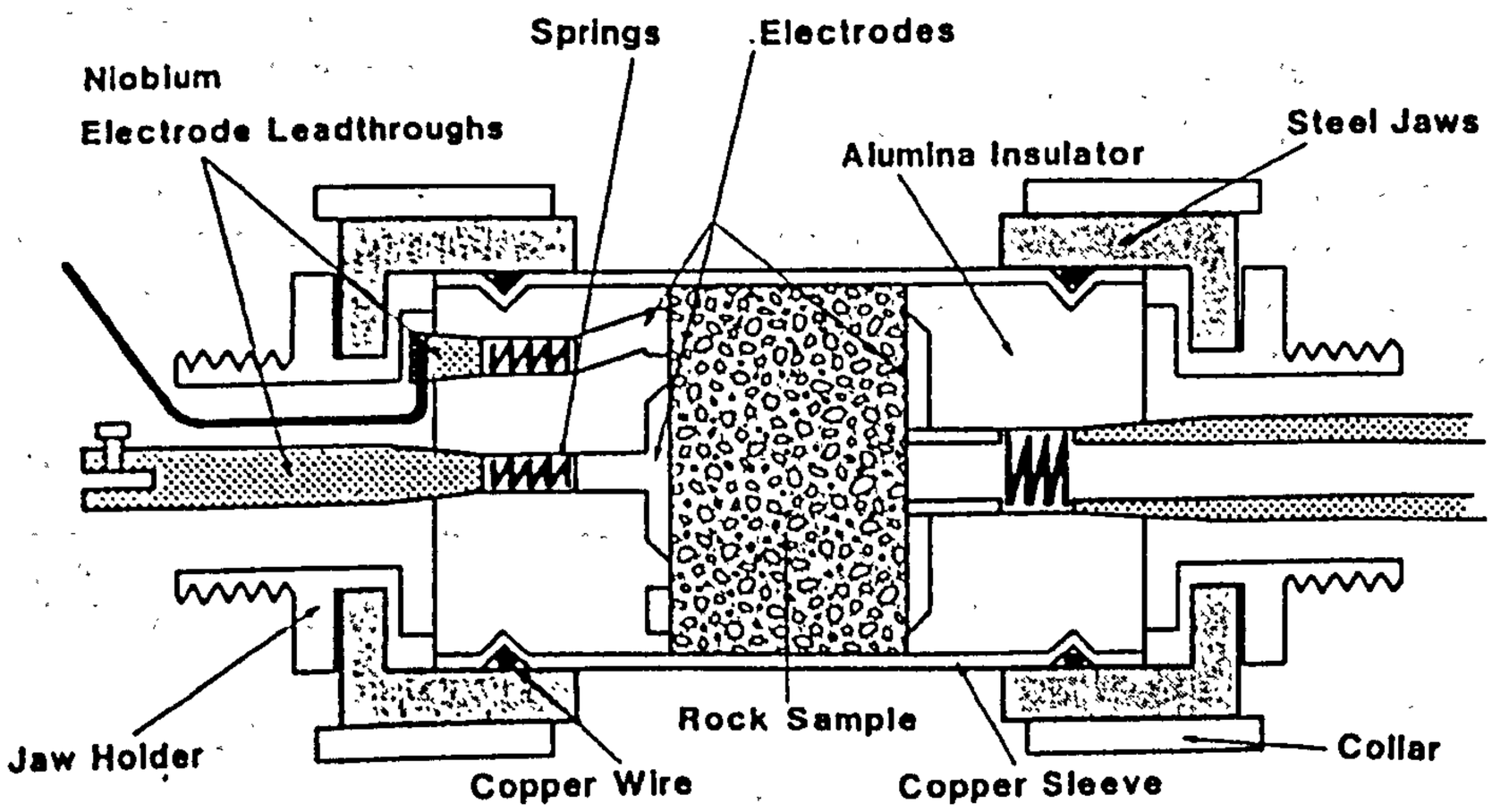

FIG. 1 The Copper Sleeved Guard-ring Experimental Cell

COPPER SLEEVE LEAKAGE CURRENTS AND THE USE OF A GUARD-RING

The electrical conductivity of the copper sleeve gives rise to large leakage currents through it. This shunts most of the current around the sample. It also distorts the current flow around the periphery of the sample near to the sleeve due to the boundary conditions that the conducting sleeve imposes on the system. The total effect of the copper sleeve is to increase the apparent conductivity over 300% compared with the unsleeved case. The problems caused by the copper sleeve can be removed by the use of a guard-ring. An annular guard-ring electrode that surrounds the collector electrode and is held at the same instantaneous potential as the collector electrode with respect to the emitter electrode, collects the leakage currents that flow through, or are distorted by the copper sleeve and drain them to ground without them being measured. The result is that the conductivity measuring bridge measures a central cylinder of rock where the flow of current is dependent upon the conductivity of the rock alone and the edge effects caused by the copper sleeve do not affect the measurement. The guard-ring potential is generated by active operational amplifier circuitry configured as a

voltage follower. It can provide a guard-ring potential that differs less than 0.1% from the collector potential, isolate the guard-ring from the collector and drain the required leakage current. This is due to its characteristics of a very high input impedance and low output impedance at the conductivity bridge operation frequency of 1592 Hz. High slew rate versions ($2x10^{+7}$ V s^{-1}) of the commonly available 741 op-amps were used to enable close potential mapping at high bridge potential outputs. Other specifications of the op-amp enable the full range of unsaturated rock conductivities to saturated rock conductivities to be measured. The measurement set up is shown in Figure 2. A Wayne-Kerr B642 Autobalance conductivity bridge has been used. The op-amp circuitry has been applied to a simple array of resistors modelling the behaviour of a loaded cell with an arranged leakage current 3.41 times the 'sample' resistor current. The measured apparent resistance of the 'sample' resistor was within 0.5% of its true resistance with the guard-ring in operation whereas it fell to less than 62% of its true value with the guard-ring inoperative. Tests of the guard-ring circuitry on a proper model of the copper sleeved cell configuration containing a saturated rock sample has given consistent results that systematically differ from the PTFE sleeved case by less than 4%. Comparing this with the 300% plus increase in apparent conductivity with the guard-ring circuitry inoperative shows how effective this method is. It is hoped that the 4% error of this technique will be able to be reduced especially if used with unsaturated rocks.

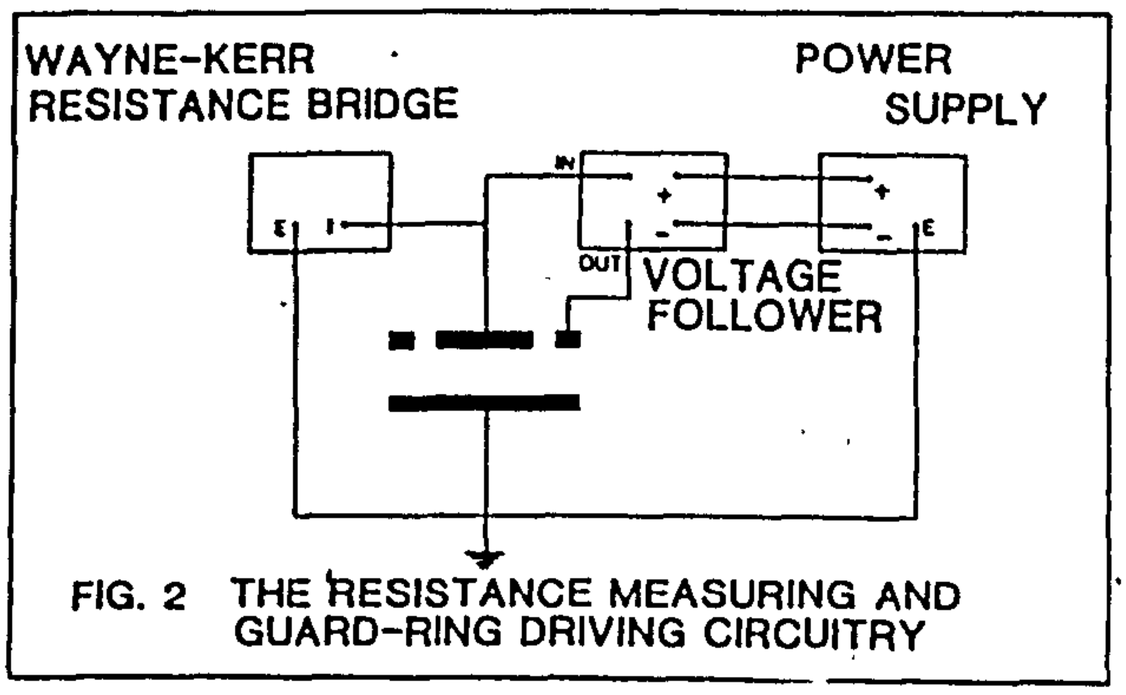

FIG. 2 THE RESISTANCE MEASURING AND GUARD-RING DRIVING CIRCUITRY

CONCLUSIONS

Results from tests on the new measurement cell and its guard-ring driving circuitry show how well they seem suited to carrying out the task for which they were designed. It is hoped that the coming year will produce saturated rock conductivity measurements at higher temperatures and pressures.

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PAPER 2:

GLOVER, P.W.J. et al., 1989
The Measurement of Saturated Rock Electrical Conductivity at Lower Crustal Temperatures and High Pressures.
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THE MEASUREMENT OF SATURATED ROCK ELECTRICAL CONDUCTIVITY AT LOWER CRUSTAL TEMPERATURES AND HIGH PRESSURES

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Laboratory measurement of the electrical conductivity of saturated rock samples has been possible at temperatures up to 900°C, confining pressures of 0.2 GPa and raised pore pressures. High temperatures constrained the sample to sit in a thin metallic sleeve. The leakage currents introduced by this method of sleeving were reduced to an acceptable level by the use of a guard-ring method. All electrodes were set into specialist ceramic holders. New seals have been developed for sealing the metallic sleeve and electrical leadthroughs to ceramic electrode holders. Experimental design will be discussed in full.

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Abstract A method has been developed to measure the electrical conductivity of electrolyte saturated rocks under simulated lower crustal temperatures and pressures. The method uses a metal sleeved sample, and a guard-ring electrode system has been used to minimise leakage currents through the sleeve. Seals were developed to mate the metallic and ceramic parts of the cell.

Keywords: Lower crust, Electrical conductivity, Guard-ring, Saturated rock, High pressure, High temperature.

1. INTRODUCTION

Field studies have revealed a zone of high conductivity in the continental crust. Several theories have been proposed to account for this; aqueous electrolyte, partial melt, conductive minerals, graphite. Non of these theories have been proven due to the lack of knowledge of how the electrical conductivity of rocks varies at lower crustal conditions. Experiments have been done on saturated rocks to upper crustal conditions (0.4 GPa and 250°C)(**Lee, 1977**) and there have been measurements on dry rocks at higher temperatures or high pressures alone. An extreme condition cell has been designed so that saturated rocks could be measured at 1 GPa confining and pore pressures, and temperatures up to 900°C to obtain data to differentiate between the various high conductivity layer hypotheses.

2. CELL DESIGN

The final cell design (figure 1) is a fully developed version of an interim design (**Glover et al., 1988**). The high specification temperatures required the cell to be designed to operate in a gas pressure vessel. The separation of gas confining pressure medium from the pore fluid and the ability to independently vary the pore pressure required the sample to sit in a sleeve. The sleeve needed to be deformable to transmit the confining pressure while being refractory to lower crustal temperatures in an inert gas environment. A thin metal sleeve was found to be the only method of fulfilling these criteria. Leakage currents arising from the metallic sleeving were 'filtered' out using guard-ring electrodes. This system requires the electrodes to be set in insulating electrode-holders at each end of the sample. There are associated

problems of sealing the electrode-holders to the metallic sleeve and leadthroughs.

3. CELL CONSTRUCTION

The electrodes were set in cold isostatically pressed pure recrystallised alumina holders. Other types of alumina were tried and rejected because of their non-zero connective porosity. The alumina fulfills the criteria of high compressive strength, high electrical resistance and stability at lower crustal temperatures. The holders have an important role to play in the cell as they hold the electrodes, have provision for the electrode connections in the form of niobium cone leadthroughs and are sealed to the sleeve. The electrodes have been made from stainless steel although platinum would improve resistance to corrosion. Machining of the ceramic was done by ultrasonic drilling and grinding with fine points. The cone leadthroughs consisted of a diamond-ground self-locking taper grommetted with copper powder. Niobium was used to ensure that the thermal expansion of the leadthrough caused minimal tensile stresses in the ceramic; niobium having a similar thermal expansivity to the ceramic. The sleeve/electrode holder seals were made by swaging the sleeve into a groove in the ceramic using a wheel tool. The seal was held in place at high temperatures by a copper circlet, four jaws and a constraining collar. Copper was initially used as a sleeving material but the reaction with pore fluid at high temperatures caused significant corrosion so 100 micron thick stainless steel sleeves were used instead. Each sleeve was used only once due to the permanence of the sleeve seal and inner surface corrosion of the stainless steel sleeves.

4. GUARD-RING ELECTRODE MEASURING SYSTEM

The presence of a metallic sleeve shunts most of the measuring current around the sample and distorts the current flow within the sample. The sleeve causes the apparent conductance of a typical saturated rock sample to rise by over 300% compared with the use of an insulating sleeve. The use of a guard-ring has removed the effect of the metal sleeve. This was done by driving an annular guard-ring electrode at the same instantaneous potential as a collector electrode. The leakage currents were collected and drained to earth while the collector gathered the current that had passed through a central cylinder of rock so it could be measured. The sample conductivity was calculated knowing this current, the potential across the sample and the geometric calibration factor of the central sampled cylinder of the rock sample. The guard-ring potential was generated by an op-amp voltage follower circuit (figure 2)(Glover et al., 1988). Measurements were made at a frequency of 10^4 rad s^{-1} on a Wayne Kerr B642 conductance bridge. The measurement circuitry was tested on an array of resistors modelling the 'leaky' sleeve scenario and enabled the 'sample' resistance to be measured to within 0.5% of its true value despite over 75% of the total current being shunted through a 'sleeve'

FIGURE 1. THE HIGH TEMPERATURE AND PRESSURE ELECTRICAL CONDUCTIVITY CELL

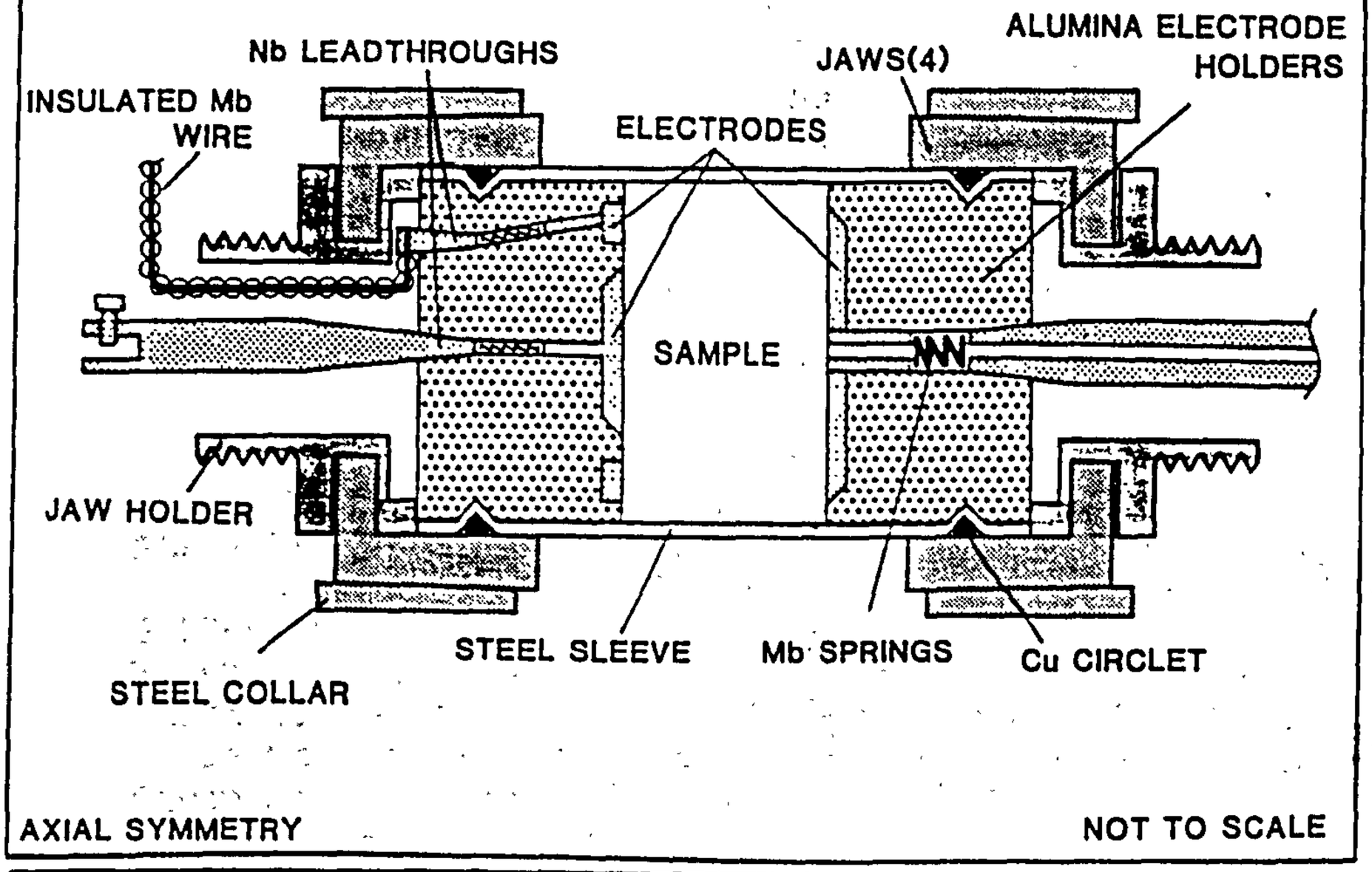
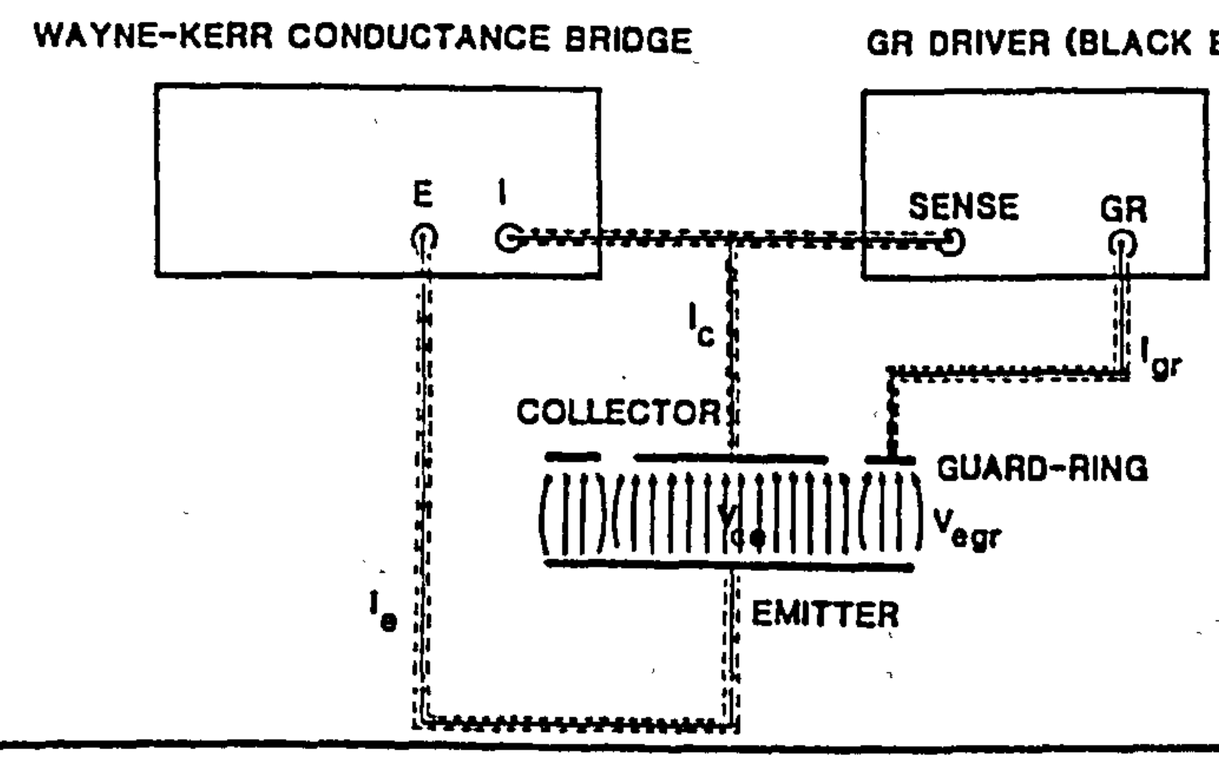


FIGURE 2. THE OPERATION OF THE GUARD-RING CONDUCTIVITY MEASUREMENT CIRCUITS



resistance. Comparisons between other methods of making conductance measurements on saturated rocks showed that this method has a systemmatic error of +4% which was eliminated on calibration to conductivity.

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MEASUREMENTS OF SATURATED BASIC AND ACIDIC ROCK ELECTRICAL CONDUCTIVITIES AT LOWER CRUSTAL TEMPERATURES AND HIGH PRESSURES

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Laboratory measurement of the electrical conductivity of saturated rock samples has been possible at lower crustal temperatures. Basic rock samples have shown increasing conductivity at temperatures above 350°C whereas acidic rocks have shown a decrease in conductivity above this temperature. This behaviour can be explained by acidic rocks relying entirely upon fluid conduction and basic rocks having additional non-fluid conduction mechanisms. The results can help explain the mid-crustal high conductivity layer and predict that fluid saturation in the lower lower crust is unimportant to its conductivity.

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MEASUREMENTS OF SATURATED BASIC AND ACIDIC ROCK ELECTRICAL CONDUCTIVITIES AT LOWER CRUSTAL TEMPERATURES AND HIGH PRESSURES

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Abstract Laboratory measurements of the electrical conductivity of brine saturated acidic and metabasic rocks were done at confining pressures up to 0.2 GPa, high pore pressures and temperatures of 900°C. Acidic rocks showed conductivities insufficient to explain the lower crustal high conductivity layer. Basic rocks, however, showed conductivities consistent with the high conductivity layer at those temperatures that the layer is thought to possess.

Keywords: Lower crust, Electrical conductivity, High conductivity layer, Saturated rock, High pressure, High temperature.

1. INTRODUCTION

Field determinations of the electrical conductivity profile of the continental crust have commonly revealed a high conductivity layer (HCL) in the upper part of the lower crust. Saturation with brine, partial melting, mineral conduction and graphite have all been invoked to explain this layer. Discrimination between these hypotheses has not been possible due to the lack of data concerning the variation of the conductivity of rocks at lower crustal conditions. Up until now data has been confined to saturated rocks under upper crustal conditions (250°C and 0.4 GPa) or to higher temperatures or pressures on dry rocks. The data presented here can be of particular use as it includes measurements of saturated samples at lower crustal temperatures, and pressures which can be extrapolated to lower crustal values.

The experimental cell used is fully described in Glover et al.(1988; 1989) and Glover (1989). 21.5 mm diameter by 27.5 mm long samples were used. The initial and repeatable conductivity variations were measured for pressures up to 0.2 GPa at low and a range of high temperatures, and temperatures to 900°C with different pore pressures and a constant confining pressure.

2. CONDUCTIVITY VARIATIONS OF SATURATED ROCKS

Figure 1 shows the electrical conductivity variation relative to that at 25 degrees C of typical saturated acidic and metabasic rocks with temperature. This variation completely dominates the variations of conductivity with pressure. The magnitude and the shape of variations of conductivity with temperature depended strongly upon the rock composition. The acidic rocks had low

conductivities and showed an increase in conductivity taking place until about 400°C followed by a rapid decrease which was faster for lower pore pressures. The conductivity of the saturated acidic rocks at lower crustal temperatures (0.0001-0.001 S m^{-1}) is small compared with that observed from field data (0.03-0.1 S m^{-1}) and such these acidic rocks cannot explain the HCL. The behaviour can be explained by an insulating rock matrix with conduction controlled entirely by the variation of the conductivity of the pore fluid with temperature. The brine conductivity variation initially increases with temperature. As temperatures are increased, however, the isobaric thermal expansion of the fluid causes a reduction in the charge carrier density and hence conductivity. The peak occurs at the temperature when the Arrhenius conductivity increase is offset by the drop due to the decrease in charge carrier density. Increase in the fluid pressure allays the reduction in the charge carrier density and hence the conductivity at high temperatures.

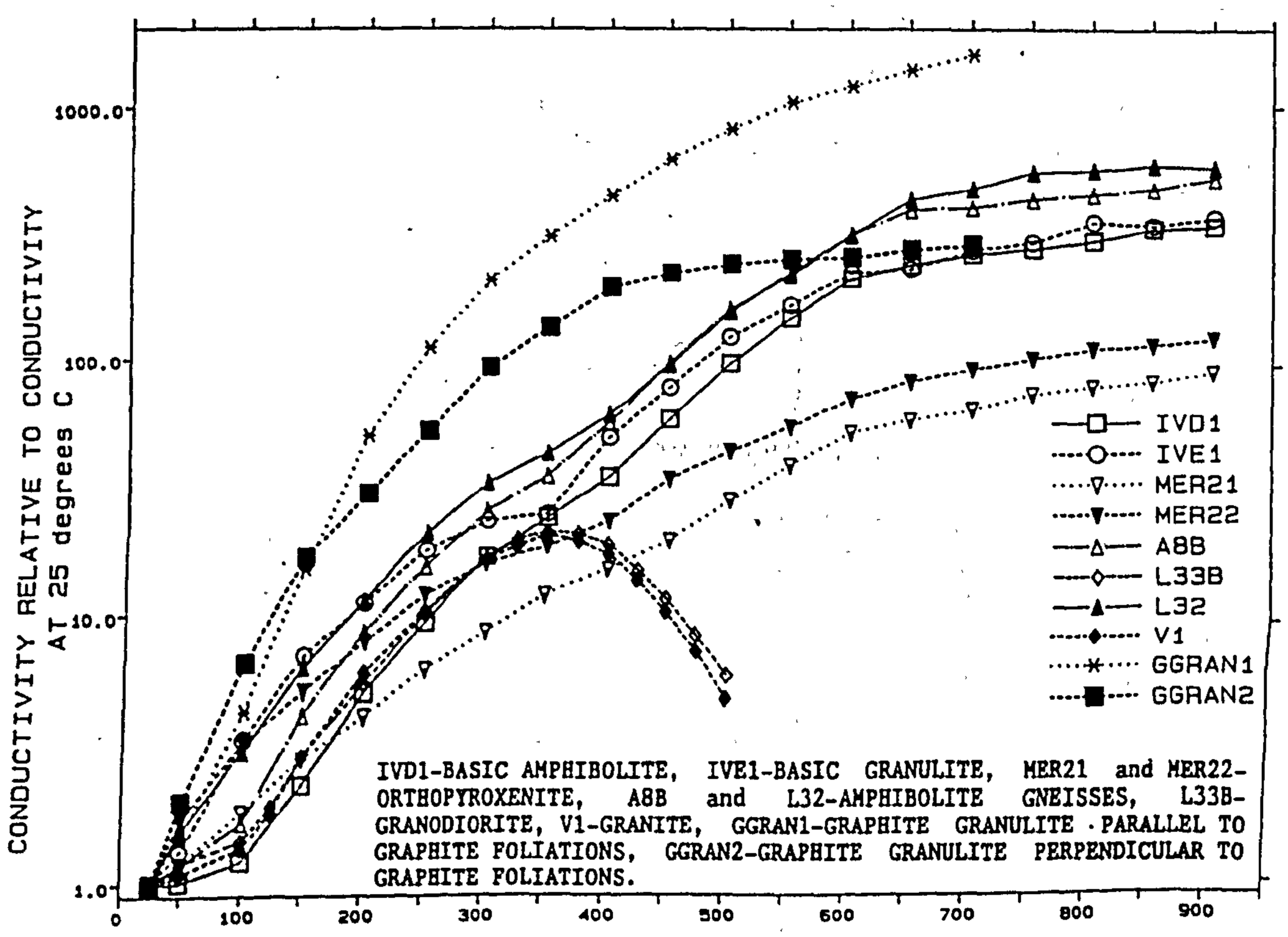
The metabasic rocks have much higher conductivities despite similar porosities. They showed a similar variation to 350°C followed by further increase in conductivity with temperature, the rate of conductivity increase diminishing after about 700°C. The magnitude of the conductivity of these types of rock at lower crustal temperatures (0.01-0.1 S m^{-1}) is consistent with the field observations for the HCL. The behaviour can be explained by electrolyte conduction plus one or several additional conduction mechanisms that increase the conductivity rapidly above 200°C. These mechanisms compensate for the loss of electrolyte conduction at all but the highest temperatures. The presence of large amounts of spinels especially magnetite and titanomagnetite, hydrous minerals, minerals with large cations and in one case graphite suggest that possible mechanisms include solid state mineral conduction.

3. CRUSTAL ELECTRICAL CONDUCTIVITY

The results primarily indicate that the lower crustal HCL cannot be due to the presence of saturated acidic rocks but must contain rocks with some conduction mechanism in addition to electrolytic conduction, such as metabasic rocks. The conductivity of the upper crust can be explained in terms of brine saturated rock with high fracture porosity. The conductivity depends upon decreases due to the increasing confining pressure and increases due to increases in temperature. The HCL can be explained by invoking the peak in the brine conduction, and the increase in conduction through magnetite-like minerals, hydrous minerals and minerals containing large cations. The temperature at levels which the HCL exists correspond to the upper part of the range where brine conduction is a maximum. It is also probable that the crust at the level of the HCL is composed of amphibolite grade facies which has a mineralogy of hydrous minerals with more spinels and minerals with large cations than most other rocks. Brine saturation is possible, with water arising from the dehydration of the amphibolite facies rocks at the amphibolite/granulite facies isograd, being generated

at high temperature shear zones in amphibolites or being carried down into the crust by deep major faults and thrusts. The porosity available for saturation at this depth has, however, been greatly reduced by confining pressure reducing the maximum conductivity ascribable to the electrolyte.

FIGURE 1. THE VARIATION OF THE CONDUCTIVITY RELATIVE TO THAT AT 25 degrees C WITH TEMPERATURE FOR DIFFERENT ACIDIC AND METABASIC ROCKS



The subsequent reduction in conductivity in the lower part of the lower crust can be explained by the loss of electrolyte conduction and hydrous mineral conduction.

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