THE GENERATION OF STORMFLOW ON A GLACIATED HILLSLOPE IN COASTAL BRITISH COLUMBIA

by

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ABSTRACT

An investigation into the mechanisms of stormflow generation on a glaciated hillslope in coastal British Columbia has been undertaken. The investigation included a controlled irrigation-runoff experiment on a 30 x 30 m hillslope plot in the U.B.C. Research Forest near Haney, B.C. Instrumentation included 12 rain gauges, 45 piezometers, and 2 outflow-tipping buckets. Piezometer slug tests to measure hydraulic conductivities and a geologic study to establish the representativeness of the experimental results were conducted to complement the irrigation experiment .

The hydrogeologic units of the research plot consist of:

A) 0.1 to 0.3 m of forest floor material consisting of organic material in various states of decay

B) 0.3 to 0.8 m of heterogenous, red-brown B horizon containing many organic rich channels made up of live and decayed roots

C) 0.5 to 2 m grey to grey-green Vashon till

D) fractured to unfractured granodiorite bedrock

The hydraulic conductivity of the till was approximately 10^{-7} m/s. A slightly higher value of 10^{-6} m/s was found for the lower B horizon matrix. A bulk conductivity for the lower B horizon was estimated at 10^{-4} m/s. The 2 to 3 order-of-magnitude difference is probably attributable to numerous, high conductivity root channels present throughout the lower B horizon.

Stormflow was generated when the water table rose into the high conductivity B horizon. Outflow at the stream bank exited from the B horizon with most water flowing from high conductivity root channels. The rate of outflow was controlled by the position of the water table. Since the water table remained parallel to the overall hillslope, the hydraulic gradient remained approximately constant. Only the cross-sectional area available for flow varied. Once outflow had commenced, the rate of outflow was sensitive to variation in the rainfall rate. Input-outflow lag-times were as little as one hour. The time lag to initiation of outflow was 19 hours. Most of this lag was attributable to the filling of storage requirements after a two month period of no rain.

The distribution of the hydrogeologic units in the research plot was found to be representative of the research area. Lag times were found to be in the range found in another similar B.C. mountain basin. It is concluded that the mechanism of stormflow generation operating in the research plot can be generalized to other similar basins.

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Many people have given me much needed advice and assistance along the way. I would like to thank them here. Jan deVries, Department of Soil Science, gave much of his time for discussion of ideas, hand-soiling fieldwork, and standing under sprinklers, piezometer measuring stick in hand. He also provided me with an on-going research project which included a research site far superior to any I could have set up myself. R. Allan Freeze, Department of Geological Science, gave me an understanding of how to do scientific research. He also amazed me at the fast turnaround on my thesis drafts. Dr. J. Leslie Smith, Department of Geologic Sciences, spent much of his all-too-precious time as a new faculty member reading and criticizing my written work. W. H. Mathews, Department of Geological Sciences, provided invaluable direction with regard to the glacial geology. R. Pamela Reid and Herman Felderhof assisted me with my fieldwork well beyond the call of friendship. C.J. Larsen gave me encouragement during and long before my graduate studies began. She also opened doors into the written world which allowed the writing of this thesis to be much less painful than it might have been. At times it was even fun! E.H. Perkins helped to de-mystify the wonders of MTS. Gordon Hodge produced beautifully drawn figures from my hasty sketches and Cathi Lowe fed my manuscript to the computer at UBC to produce this printed copy. Research funds were provided by the National Research Council of Canada.

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1.0 INTRODUCTION AND RESEARCH OBJECTIVES

Throughout history many people have speculated on the origin of rivers and streams. However, it has only been in the last 50 years that attempts have been made to explain the mechanisms of stormflow generation. Basically, three mechanisms have been proposed: overland flow caused by rainfall in excess of infiltrability, subsurface stormflow and overland flow caused by a rising water table. All three mechanisms are known to operate under certain circumstances but it is not clear which one predominates in the mountains in coastal British Columbia.

Recent work in the south-west Coast Mountains of British Columbia, has shown that extreme soil heterogeneity may favor a subsurface mechanism of channelized flow through organic zones. DeVries and Chow (1973, 1978) have shown that soil heterogeneity plays a major role in infiltration in the Seymour Watershed. Continuing this work, Nagpal and deVries (1976) established an experimental hillslope-stream system in the UBC Research Forest in order to better understand stream flow in forested mountain environments. Their findings can be summarized as follows:

1) stormflow traveled through the hillslope via root channels, by-passing the soil matrix
2) outflow from the experimental watershed decreased as water broke through imperfections in the till which underlay the soil
3) leakage through this till was up to 75% of the total rainfall input

The current study reports on work carried out by the author and deVries to critically examine these conclusions.

One objective of this study was to examine the underlying glacial till with emphasis placed on the hydrologic behavior and the spatial distribution both within and near the research plot.

Another objective of this study was to investigate the mechanisms of stormflow generation within the experimental hillslope-stream system and to examine the role of root channels in the B horizon of the soil. Previous work by Nagpal and deVries (1976) had shown that these channels were major conductors of stormflow to the stream bank. However, because this mechanism of stormflow generation aroused some controversy, it was decided to re-examine these conclusions using more complete instrumentation.

Finally, it was important to establish the generality of the flow mechanisms by examining and comparing the geology in the research plot to that of the area around it.

The plot hydrology was examined using a controlled irrigation experiment on the same site used by Nagpal and deVries (1976). In order to positively establish steady -state conditions and to better understand flow paths, more complete instrumentation, chemical tracers and a longer period of irrigation were used.

This thesis reviews the literature and presents the theory of various mechanisms of stormflow generation. The local geology is then explored using previous work and that done by the author to show the hydrogeologic representativeness of the experimental site. The irrigation experiment and the related instrumentation are then described followed by the results and analysis.

Finally, conclusions are presented along with discussion about generalization of the results to other areas.

This study does not answer all the questions concerning stormflow generation in the coastal mountains of British Columbia. However, it is hoped that through this study the reader will gain a better understanding of mountain forest hydrology.

2.0 LITERATURE REVIEW

In this chapter three mechanisms for stormflow are reviewed: Hortonian overland flow, subsurface stormflow, and Dunne and Black overland flow. These are discussed chronologically as done by Freeze (1974) and Engman (1974). Recent work in the southwest coastal region of British Columbia on stormflow generation is then summarized.

2.1 Hortonian Overland Flow

The mechanism of stormflow generation by overland flow was explained by Horton (1933). In his now classical interpretation of the role of infiltration, he stated that precipitation in excess of that which could infiltrate into the ground, fills surface depressions and then flows along the surface to the stream channel in the form of overland flow. This "precipitation excess" occurs whenever the rainfall rate exceeds the soil's maximum possible infiltration rate, or infiltration capacity as Horton called it.

Horton stated that the infiltration capacity, or infiltrability as it is now often termed, is not constant but decreases with time during a rainfall event to an approximately constant rate. It then returns to its initial value within a few days after precipitation has ceased. According to Horton, the decrease in infiltrability is due mainly to three factors: packing of the soil surface by rain drop impact, swelling of the soil and closing of surface openings, and inwashing and filling of soil surface openings by fine materials. It was also stated that the subsequent return to the pre-storm infiltrability is caused by a reversal of the above by sun, wind, and biologic action on the surface of the soil.

Horton noted that infiltrability varies with soil type and the time of year during which rain occurs. Fine textured soils have infiltrabilities that decrease more rapidly and level off to lower rates than coarser soils. He also noted that maximum infiltrabilities occur during summer months when higher temperatures and more active biota produce a greater degree of surface restoration.

As Horton viewed it, overland flow occurs whenever rainfall rate exceeds infiltrabilities, as shown in Figure 2-1. This type of mechanism is dominant where rainfall rates are frequently higher than natural soil infiltrabilities (arid to semi-arid regions with thunderstorm activities) or in places where the soil surface is disturbed (agricultural land or urbanized areas). These regions commonly have exposed soil surfaces and little vegetation for protection from raindrop impact. The lack of vegetation also reduces the organic content of the soil and therefore reduces the hydraulic conductivity, too.

Horton claimed that the decrease in infiltrability with time is due to effects on the surface of the soil and not due to the effects of saturation. Although experimental evidence by Green and Ampt (1911) showed a decrease in infiltration independent of surficial effects, Horton did not take this into account. His work was empirical, a result of many observations but with no physical basis for his theory. A theoretical understanding came later.

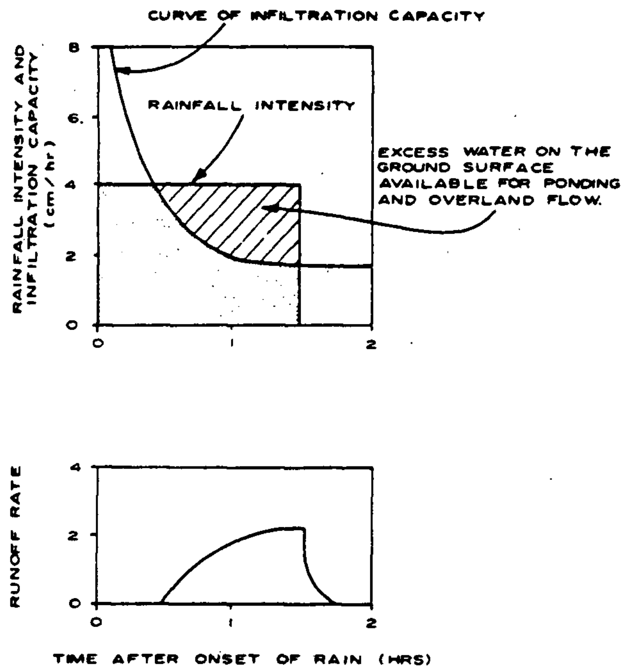

FIGURE 2-1 Rainfall, Infiltration and Hortonian Overland Flow (after Freeze, 1974)

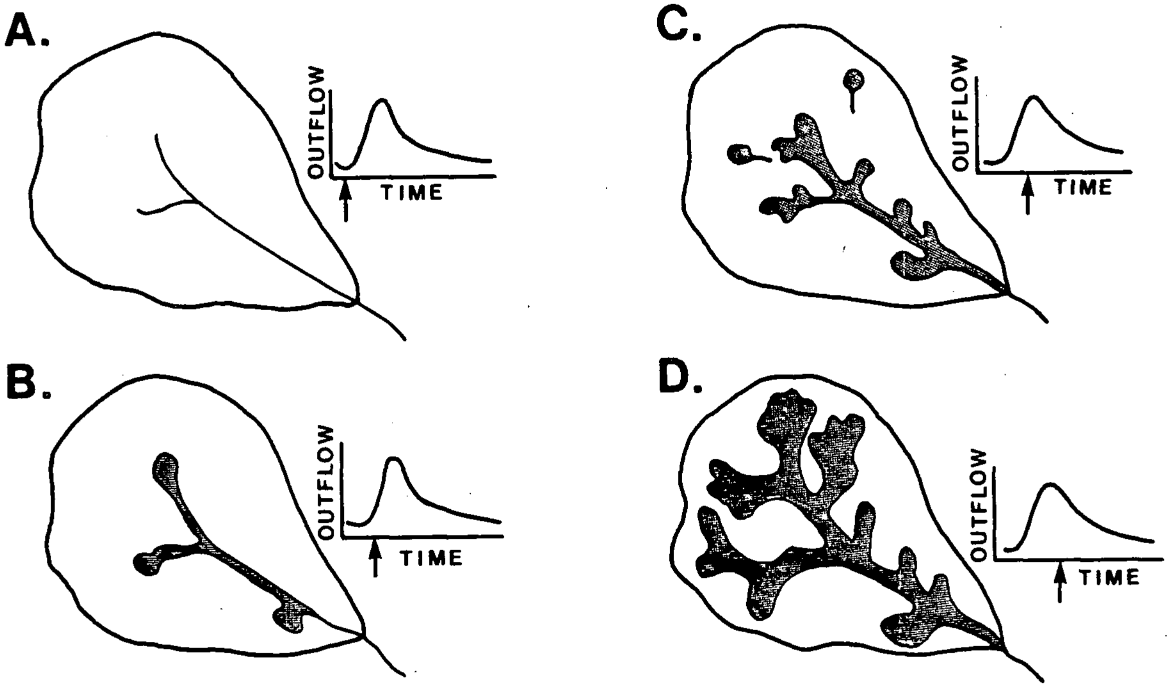

FIGURE 2-2 The Subsurface Stormflow Contributing Areas of Hewlett and Nutter (after Hewlett and Nutter, 1970)

While Green and Ampt (1911) provided a semi-theoretical explanation for infiltration, it was not until the work of Rubin and his co-workers in the 1960s that a complete theoretical understanding was produced. Richards (1931) developed the equation of unsaturated flow based upon Darcy's law and the equation of continuity and Philip (1957) provided the first analytical solution. Numerical solutions by Klute (1952), Day and Luthin (1956), and Hanks and Bowers (1962) helped lead to the lucid understanding provided by Rubin and Steinhardt (1963) and Rubin et al. (1964).

The work of Rubin and his co-workers revealed that surface ponding occurs only if rain falls at a rate greater than the saturated conductivity of the soil and if it does so long enough for surface saturation to take place. Thus, overland flow is produced whenever saturation occurs at the surface. It is not just a function of surface effects but also of initial soil moisture and the soil properties that affect the soil's unsaturated response to wetting.

Horton's theories implied that rainfall usually exceeds infiltrability and that stormflow is typically produced by overland flow. He inferred that such overland flow generation is areally widespread. However, within a watershed, soils usually have considerable heterogeneity and rainfall can vary both temporally and spatially. Recognizing this, Betson (1964) and TVA (1965) developed a partial area concept whereby only certain areas, usually with less vegetation and higher soil moisture contents, consistantly produce runoff to streams in the form of overland flow. This partial area concept helped to explain why most watersheds in humid areas typically generate storm runoff of less than 10% of the total rainfall input.

Subsequent work by Whipkey (1965), Ragan (1968), Dunne and Black (1970a, b), Weyman (1970), and others has shown that rainfall rates in temperate, middle latitude areas rarely exceed infiltrabilities. Thus, Hortonian overland flow is rarely seen in these environments. This is particularly true in forested regions where dense vegetation protects the soil surface and where conductivities are high because of coarse texture and high organic content. It is not surprising that workers in these areas proposed a new mechanism for the generation of stormflow.

2.2 Subsurface Stormflow

In vegetated regions with less intense rainfall and permeable soils, Hortonian overland flow is rarely seen. The lack of Hortonian flow, coupled with observations that streams in these areas do respond to precipitation, led to the development of the second major concept of stormflow generation: subsurface stormflow. This is a mechanism whereby storm water flows to the stream channel via shallow subsurface paths. It requires both steep hill slopes and large hydraulic conductivities in a shallow soil horizon (Freeze, 1972), a situation commonly found in upland forests

Although originally postulated by Hursh (1936, 1944) it was not until Hewlett and his co-workers (Hewlett, 1961; Hewlett and Hibbert, 1963, 1967; and Hewlett and Nutter, 1970) that this mechanism became favored by forest hydrologists. Hewlett (1961) and Hewlett and Hibbert (1963) established that soil could provide baseflow between rainfall events via coupled saturated-unsaturated flow. After thoroughly watering a 1 x 1 x 14 m soil block and covering it to exclude atmospheric interaction, they measured an outflow of 1.42 liters per day after 60 days. This was admittedly small but it did demonstrate the feasibility of downslope movement of water through unsaturated soil.

Continuing this work, Hewlett and Hibbert (1967) proposed a mechanism where areas nearer the stream bank play a more significant role in stormflow generation than areas farther away. Based on the partial area concept of Betson (1964) and TVA (1965) and the "translatory flow" ideas of Horton and Hawkins (1965), they proposed a variable source mechanism of stormflow generation. In this mechanism, runoff is produced by subsurface stormflow instead of Hortonian overland flow and contributing areas are not static but expand with increasing amounts of rainfall during a storm. According to Hewlett and Hibbert, the expansion of these source areas to meet the subsurface flow paths in the hillslope allows the relatively slow moving soil water to reach the stream quickly enough to account for rapid stream rises (Figure 2-2).

These low subsurface flow velocities and the short observed rainfall-streamrise lag-times are a basic problem with the subsurface stormflow concept. Field measurements by Whipkey (1965) on a layered 1 x 2 meter plot on a 28% hillslope in Ohio, using 24 simulated rainfall events with both wet and dry antecedent soil moisture conditions indicated minimum inflow-outflow lag-times of 1 1/2 hours. Other "storms" of lower and more typical rainfall intensities yielded lags that were longer.

Initiation of outflow came as late as 2 1/2 hours after the beginning of rainfall. Extrapolation of this small soil block to watershed dimensions would not correlate well with actual stream response.

Experimental work by Weyman (1970) in Somerset, England yielded similar conclusions. On a one meter wide hillslope plot (length not given), he noted hillslope lags of 36 hours for subsurface stormflow as opposed to lags of 3 to 4 hours for the actual stream peak. Weyman's explanation for this was an unmeasured and more favorable portion of the basin upstream from his site. Both Whipkey and Weyman had long lag-times and hillslope flow volumes that were too small to account for stream rises. More information on subsurface stormflow was needed to explain what was actually happening.

This explanation was provided by a careful field study by Dunne and Black (1970a, b). In a well instrumented basin in Vermont (discussed in more detail in the next section), they concluded that subsurface stormflow was only seen on steep, laterally concave slopes with wet antecednt conditions and intense rainfall. Even with these most favorable conditions, they measured total volumes that were too small and lag-times that were too long to produce the rise seen in their stream.

These results were verified by Freeze (1972a,b) with computer simulation of a small watershed. By using finite difference approximations for the differential equations of flow, he was able to simulate a small watershed by coupling boundary conditions for the hillslope surface, subsurface and stream channel. By varying initial conditions, hillslope parameters and boundary conditions, he was able to simulate a wide variety of watersheds, including those thought most favorable for subsurface stormflow. He found that subsurface stormflow is a major peak contributor only with longitudinally convex hillslopes with highly permeable soils feeding steeply incised channels.

These studies indicated that subsurface stormflow is relatively unimportant in regions with homogeneous soils. However, evidence exists that in heterogenous systems, subsurface flow is important. Whipkey (1967), using many (total not given) hillslope plots of up to 1,100 m^2 on slopes of 19 to 42% with rainfalls of 12 to 76 mm per hour, stated that in finer soils, subsurface flow travels via biological and structural channels. His tensiometers indicated a by-passing of the soil matrix. Outflow lag-times were as low as 15 to 25 minutes even with flow paths that passed through 1.22 m of unwatered buffer strips. Visual observations at the outlet of the experimental sites indicated that water flowed via root holes. In one case, he observed water flowing in channels 9 m obliquely downslope from the irrigated area. Outflow occurred within 45 minutes of surface ponding on the wetted slope. A short lag time such as this indicates that subsurface stormflow could be important in heterogenous soils.

Other examples of the importance of heterogeneity in soils include work by Pond (1971) who stated that natural subsurface flow "pipes" can be detected during dry weather by slight surface depressions and specific vegetation or located just after a storm by "listening to the water gurgling beneath the surface!" Chamberlin (1972), deVries and Chow (1973,1978), and Nagpal and deVries (1976) also demonstrated the importance of subsurface stormflow in heterogenious soils. This work will be discussed in detail in a later section.

With the exception of areas with suitable soil heterogeneities, it can be concluded that where rainfall is moderate and slopes are vegetated, neither subsurface stormflow nor Hortonian overland flow is the dominant runoff-generating mechanism. To explain runoff in these regions, a third type of flow mechanism was needed. This explanation was provided by Dunne and Black (1970a, b).

2.3 Dunne and Black Overland Flow

Dunne and Black (1970a,b) chose a small sub-basin of the Sleepers River Experimental Watershed in Vermont to look at all the hydrologic components of a small watershed. Their original objective was to examine subsurface stormflow. To do this, they instrumented three segments of a hillslope (concave, convex, and planar) that consisted of an organic rich sandy soil overlying a low conductivity lacustrine clay (Figure 2-3a). The slope was steep, 30 to 100%, and the hydraulic conductivity of the soil was high. These conditions were thought to be ideal to obtain quantitative measurements of volumes and lag-times of hillslope output. They constructed a 75 m long collection trench to observe surface, shallow subsurface and ground water flow (Figure 2-3b). On the hillslope they had 30 rain gauges, 12 piezometers and 9 neutron probe access tubes as well as 2 stream gauging weirs (Figure 2-3c). A sprinkler system allowed them to

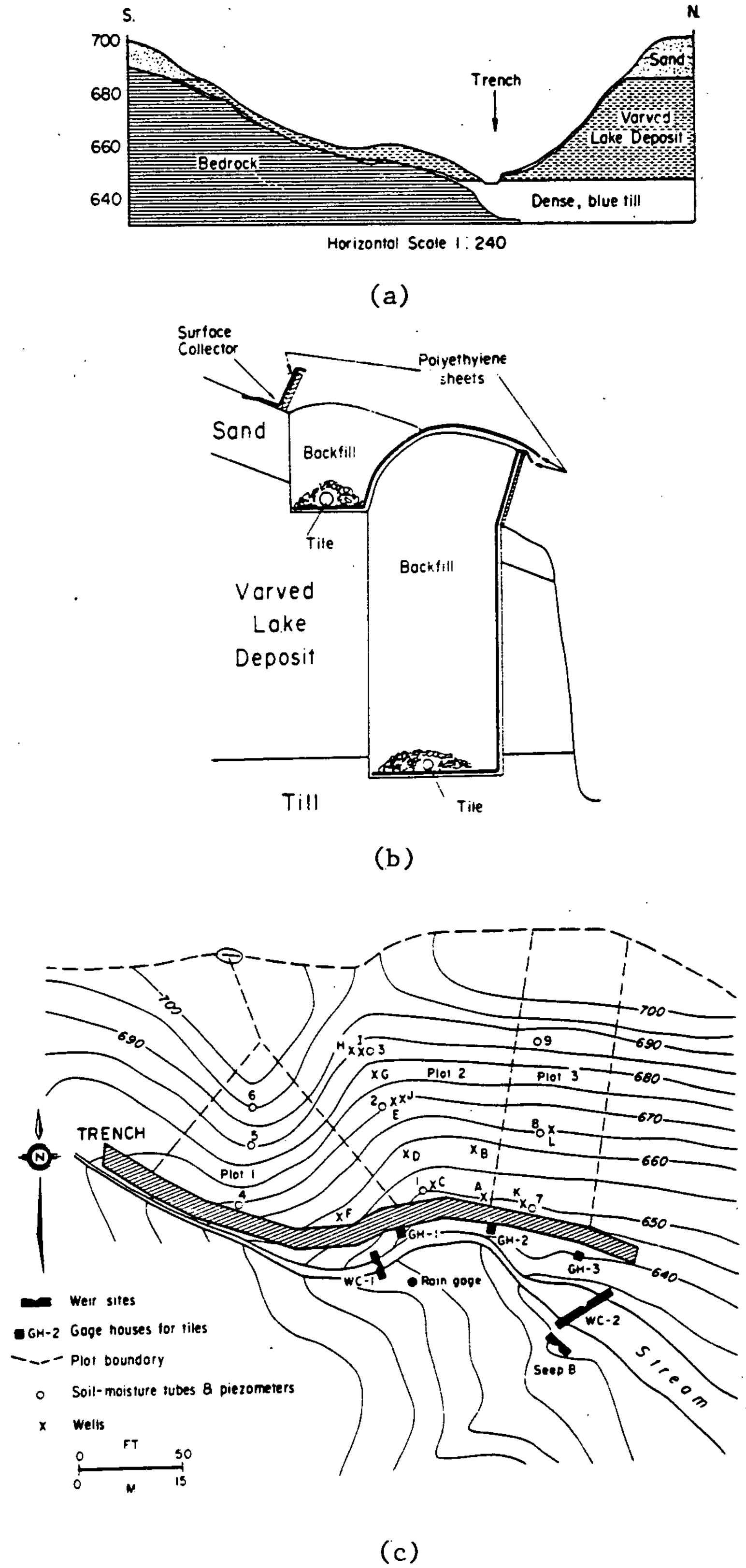

FIGURE 2-3 The Experimental Site of Dunne and Black (after Dunne and Black, 1970a,b)

produce artificial rain.

Their results were surprising. In 35 natural events with maximum recurrence intervals of 2 years, during late summer (dry antecedent moisture conditions), no significant output from the hillslope was produced. They concluded that precipitation went to replenish soil moisture lost through summer evapotranspirative demands and through ground water input. Hortonian flow was not seen as the most intense rainfall of 3.12 inches per hour was less than the measured infiltrabilitiy of 3.15 inches per hour. Later in October, conditions were wetter and, although not sufficient to create Hortonian overland flow, they were thought more likely to produce subsurface stormflow. In spite of this, runoff was negligible. Only with a large artificial storm on wet soil did stormflow appear. It was neither subsurface stormflow nor Hortonian overland flow, however.

Dunne and Black observed a type of overland flow generated on areas saturated by a rising water table. They proposed a flow mechanism that had some similarities to previous stormflow concepts. Like Hortonian overland flow under the partial area concept of Betson (1964) and TVA (1965), it too, was generated on small portions of the watershed only. These areas, though, were more like the variable source areas of Hewlett and Hibbert (1967) and Hewlett and Nutter (1970). They were generally topographically low with near surface water tables and higher antecedent moisture content. Also consistent with this concept, these saturated wetlands expanded and contracted with varying amounts of precipitation. Thus Figure 2-2, used to illustrate variable source areas for subsurface stormflow, can also be used to depict runoff producing areas like the ones observed by Dunne and Black. Here, instead of expanding channel areas fed by subsurface stormflow, the near stream contributing areas generate overland flow fed by rainfall on saturated surfaces created by rising water tables.

Like Hortonian overland flow, the Dunne and Black mechanism produces short lag-times and relatively high runoff volumes. There is, however, a major difference between Hortonian and Dunne and Black overland flow. Hortonian flow is produced when saturation occurs at the surface because rainfall rate exceeds infiltrability, while Dunne and Black overland flow is generated when saturation occurs from below as the water table rises to the surface in response to infiltration. Subsurface conditions, therefore, do not produce much runoff directly, but are important in that antecedent moisture and depth to saturation influence the extent of the water table rise.

Dunne and Black runoff comes from both direct precipitation on a saturated zone and from the saturated subsurface discharge to this zone, termed return flow. Although Dunne and Black equated the quantity of overland flow directly with the amount of rainfall on the expanding wetlands and not to the amount of return flow, an isotopic study on four basins in Canada by Fritz et al. (1976) indicated that ground water discharge may play a more important role in stormflow generation. Their study did not involve any hillslope sampling, however. Freeze (1972a,b;1974) demonstrated that return flow is dominant only when soil conductivities are high, the hillslope is steep, convex, and feeds an incised stream.

The findings of Dunne and Black were confirmed by Freeze (1972a, b) using the computer simulation technique discussed in the previous section. He concluded that in watersheds of the type studied by Dunne and Black, the flow mechanism which they described was the norm. He stated that only in extreme cases would either Hortonian overland flow or subsurface stormflow play a dominant role on humid, vegetated slopes.

Freeze's simulations were of simplified, homogeneous hillslopes. Snyder (1973), Knisel (1973) and Hewlett (1974) felt that these were inadequate to describe stormflow in some regions. In places where layering or extreme heterogeneity exist, wetting fronts and water table rises, may not be vertically continuous. Flowpaths will not be as simple as those shown by Freeze and may instead be very complex. One such area is the mountainous region of southwest coastal British Columbia. Here exists a situation that cannot be represented by a simple two layer, continuous model. This is one place where subsurface mechanisms other than those described by Dunne and Black may apply.

2.4 Work in the Southwest Coast Mountains of British Columbia

Recent work in the southwest Coast Mountains of British Columbia by Chamberlin (1972), deVries and Chow (1973) and Nagpal and deVries (1976) has shown that soil-water flow paths are much more complex than those seen in regions with more homogeneous soils. Instrumented plot studies indicated that infiltration does not progress as a continuous wetting front.

Rather, lower layers of the soil sometimes appear to wet-up before layers above them. The mechanism put forward by the above authors to explain this behavior is one of concentrated saturated flow.

Chamberlin (1972) was the first to report this type of behavior which he observed in the Seymour Watershed near Vancouver, B.C. He conducted several irrigation experiments on an instrumented 4 m^2 plot on a hillslope of 30%. The plot contained 1 m of soil over quartzdiorite bedrock. The soil contained forest floor material 0.1 to 0.3 m thick with cavities up to 1 m across under roots and stumps. The soil had a well-developed Ae horizon and a B horizon up to 1 m thick. Roots and woody material made up as much as 50% of the upper 0.2 to 0.5 m of the soil with live roots extending throughout the B horizon and along weathered bedrock surfaces. His instrumentation consisted of 5 nests of 4 tensiometers.

His results were unexpected. Tensiometers at lower levels often responded to irrigation before those located at a higher level. Some of these also indicated saturated conditions. Chamberlin's explanation for this behavior was that free water drains to lower levels in the soil via interconnected pathways that bypass the soil matrix. He did not identify the nature of these pathways but the implication was that they consisted of live and decayed tree roots.

Chamberlin suggested that these pathways are connected to the surface and, when free water is available, either from direct rainfall at the surface or because of concentration by buried logs or rocks, channelized, saturated flow occurs. This happens even though flow rates and soil parameters indicate that unsaturated conditions should prevail.

Chamberlin called a soil with many pathways connected to each other and the atmosphere, an open soil. In regions where such soils predominate, the implications for stormflow generation are considerable. Chamberlin suggested that saturated subsurface stormflow through "anomalous zones" in open soils explains the very flashy response of coastal streams to precipitation. With this mechanism, a subsurface stormflow concept different from Hewlett and Hibbert (1967) can be envisioned. The problem of the low velocities of unsaturated flow is eliminated because subsurface flow would be channelized and saturated, and therefore, much faster. Since Chamberlin did not go into detail on the complete mechanisms of stormflow generation, further research was needed to examine this hypothesis.

This research was begun by deVries and Chow (1973, 1978) using an experimental set-up similar to Chamberlin's. Their results were comparable. On a 2.5 by 3.5 m hillslope plot with 25° to 30° slopes at the 300 m elevation in the Seymour Watershed, they installed 13 tensiometers in 3 nests. The soil system was similar to that studied by Chamberlin except that the parent unit underlying the soil was a low-conductivity glacial till. High intensity irrigation (2.6 cm per hour) was applied with the soil in three states of alteration: undisturbed, partially disturbed forest floor and forest floor removed. They concluded that water moved downward primarily via root channels during infiltration but, after rainfall stopped, drained through the soil matrix.

Because of extreme soil heterogeneity, infiltrated water did not travel through the soil matrix as unsaturated flow; rather, it traveled as channelized, saturated flow. As in Chamberlin's study, this behavior was indicated by lower tensiometers that responded before, and sometime with greater magnitude than those nearer their surface. Some of these piezometers also indicated saturation, even though the rainfall rate and the soil parameters would suggest that saturation would be unlikely. They found this behavior puzzling. They stated that this response was probably not due to air entrapment nor due to pressure build up from output impediment. The implication was that localized saturation occurred.

By generating two-dimensional hydraulic potential maps from the 3 tensiometer nests with a distance weighted computer extrapolation programme, deVries and Chow analyzed the complexity of flow paths. They concluded that there was a much greater variation in hydraulic potential during infiltration than drainage. This variation indicated that root channels conducted water during infiltration but, because of reverse potential gradients, not during drainage. Free water did not enter these channels from the matrix, however. It entered at or near the surface during infiltration and only because wood, rocks, and other objects concentrated flow such that free water was available. The free water then drained into the openings of these high conductivity zones which were at atmospheric pressure.

By disturbing the forest floor, the variability in the distribution of hydraulic potential decreased. This was because channel openings near the surface were closed, eliminating the access of free water. In these "closed soils", flow was primarily through the matrix. Supportive evidence was found by Cheng (1976) who used paired watersheds to observe the effects before and after clearcut logging. He found that storm peaks were delayed and reduced on disturbed hillslopes, thus tending to confirm the importance of root channels as flow paths. This is in disagreement with Plamondon et al. (1972) who claimed that the forest floor does not effect stormflow peaks.

Continuing mountain watershed research, Nagpal and deVries (1976) studied a 30 x 30 m hillslope plot in the U.B.C. Research Forest, near Haney, B.C. to examine flow paths on a larger scale. Their instrumentation (Figure 2-4) included 7 piezometers, 4 neutron access tubes, 2 tensiometer nests and 2 calibrated tipping buckets to measure input and outflow. Artificial storms were created with 8 metered sprinklers. The soil at the research site was similar to that in the Seymour Watershed with 0.05 to 0.2 m of forest floor material, a thin layer of Ae horizon, approximately 1 m of B horizon on top of about 1 m or more of glacial till, all underlain by intrusive bedrock.

This experimental plot is the same one studied by the author and reported on in this thesis. A more detailed description appears in the following section.

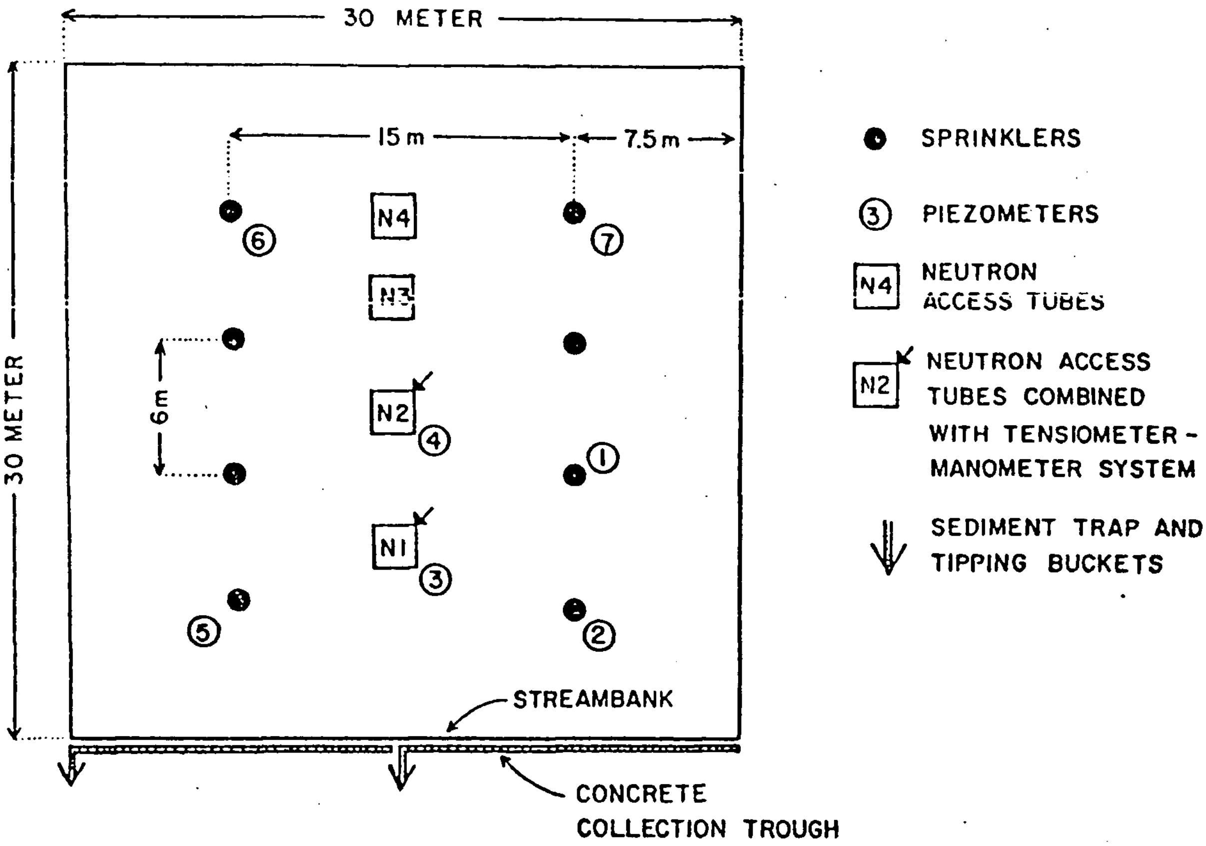

FIGURE 2-4 The Study Site of Nagpal and deVries (after Nagpal and deVries, 1976)

Nagpal and deVries concluded that waterflow from the soil to the stream bank was through root channels. This was indicated in several ways: tensiometer-neutron probe data, stream bank observations and concentration time calculations.

Tensiometers again indicated that lower parts of the soil responded before some upper layers. Also noted again were positive tensiometers readings. However, neutron probe data, as well as water retension characteristics of the soil, indicated that the soil was unsaturated. They found this puzzling. Their explanation for this response was that laboratory-measured calibration curves may not have indicated true field conditions. Another explanation seen by this author is that saturated flow occurred in the channels while the surrounding matrix was unsaturated. The neutron probe integrated over both and indicated an average unsaturated condition.

Observations at the stream bank gave visual confirmation to the importance of root channels. During full flood, a major proportion of the outflow was seen to eminate from root channels discharging at the bank. One such channel was measured at 2% of the total outflow. Also, during the initial phases of flooding, output from a distinct group of roots produced most of the total outflow for the whole stream bank.

Nagpal and deVries made a "time of concentration calculation" to deduce if initial response times were consistent with root channel flow. This calculation was based on the assumption that subsurface flow is analogous to surface flow in that outflow timing and volume are direct functions of path length and that outflow represents the water actually applied as rain rather than water from storage. They examined the time required for water to flow along the longest path length in their watershed. Using the timing of the rise to peak outflow and the length of the hillslope, they calculated a series of "times of concentration" as a function of possible matrix conductivities. This was done using Darcy's law and ignoring storage and vertical flow times. They concluded that the conductivity required for the observed lag-times was too large to represent the matrix alone and therefore indicated that a large proportion of the total flow must be conducted by root channels.

Nagpal and deVries drew several other conclusions. One was that up to 75% of the input water was lost through leakage out of the plot. Based on observations of boulders protruding through the till, surrounded by small rock fragments, they postulated that this leakage was due to water "breaking through imperfections in the compacted till." In addition, they concluded that the piezometric data indicated a discontinuous water table. This was also demonstrated in another irrigation run with 20 piezometers from which they concluded that a water table on top of the till exists but that it is only locally permanent. Other areas are saturated only under wet conditions (deVries, pers. comm.). They also noted that, unlike classical homogeneous hillslopes with higher moisture conditions nearer the stream bank, the observed water table rise was not a function of distance from the stream. On the contrary, the first and highest piezometer rise occurred near the the extreme upslope position while one piezometer near the bank did not respond at all. This led Nagpal and deVries to postulate a stream flow mechanism where topographic highs on the upper surface of the till act as contributing areas to lower, permanently-saturated areas. During rainfall, these saturated subsurface depressions fill up and overflow to the stream bank via a permeable root mat and network of root channels. These high conductivity zones allow for rapid flow even though the soil matrix conductivity is low. Thus, they postulated that the time lag betwen the initiation of rainfall and the beginning of outflow is due to the filling of subsurface basins as well as soil moisture recharge.

This is a new concept of subsurface stormflow that is not yet completely understood. Several problems remain to be examined. The investigation of these problems is the objective of this thesis. Emphasis will be placed on the glacial till, its distribution and hydroogic properties, including leakage from the hillslope system; the examination of the mechanisms of stormflow generation with emphasis on the role of the organic zones; and the geology of the hillslope plot and environs to determine the generality of any stormflow generating mechanism operating in the research plot.

To accomplish these ends, an irrigation experiment similar to those of Nagpal and deVries was undertaken during the summer of 1977. In addition to this, geologic mapping and auxillary hydrologic work were conducted.

3.0 The Study Area and Site

In this chapter the physical characteristics of the research area and experimental plot are discussed. Previous pedological and geological work in the U.B.C. Research Forest is reviewed. Then, in order to establish the representativeness of the plot, an investigation of the hydrogeological characteristics of the study plot and the surrounding area is presented.

3.1 The Study Area

3.1.1 Location and Physiography

The experimental work was carried out in the University of British Columbia Research Forest, located on the southern edge of the southwest Coast Mountains, 45 km east of Vancouver , near Haney, B.C. (Figure 3-1). The climate is Pacific Marine Humid with average daily temperatures of somewhat over 15°C for the summer months to just below 0°C during the winter. Precipitation ranges from 2.0 to 3.0 m per year with the bulk produced by Pacific frontal systems during the fall-winter-spring months. (Rainfall is not unknown, however, during July and August.) Less than 15% of the total precipitation occurs as snow because of the moderatng effects of the Pacific and the relatively low elevation. The study area is around 350 m above sea level.

Krajina (1969) places the study area in the Coastal Western Hemlock Biogeoclimatic Zone. Climax growth is Western Hemlock and Western Red Cedar. Alder predominates in recently logged or burned areas with Douglas-fir being the transitional species. Underbrush includes vine maple, brachen fern, sword fern, blueberry, huckleberry, and salal and is very dense, especially in recently cleared areas.

3.1.2 Previous Descriptions: Geology and Pedology

The study area is typically underlain by intrusive bedrock covered by thin and discontinuous unconsolidated deposits of glacial origin. The bedrock geology was mapped by Roddick (1955,1965). The surficial geology was originally discussed by Armstrong (1957) with updates and revisions in the regional quarternary stratigraphy by Armstrong (1975), Armstrong and Hickock (1975) and Hicock(1976). Soils in the Research Forest have been mapped extensively by Klinka (1976) and analyzed at the study plot by Nagpal and deVries (1976) and Bryck (1977).

Bedrock consists mainly of Cretaceous quartzdiorite to granodiorite which belong to the Coast Crystaline complex. These are intruded locally by minor andesite-basalt dykes. Because of rapid uplift and recent glaciation, exposed rock surfaces are generally fresh with little weathering.

According to Armstrong(1975), Armstrong and Hicock(1975) and Hicock (1976), the south west coastal region has undergone three major glaciations: the late Wisconsin (11,000 to 20,000 Y.B.P.), the middle Wisconsin (42,500 t0 52,000 Y.B.P.) and the early Wisconsin? or pre-Wisconsin? (> 62,000 Y.B.P.). Drift from the older stades are the Semiahmoo and West Lynn. Deposits from the late Wisconsin Fraser stade can be divided into three groups: the Sumas drift (10,000 to 11,000 Y.B.P.), the Capilano sediments (11,000 13,000 Y.B.P.) and the Vashon drift (13,000 to 20,000 Y.B.P.). Evidence exists for at least three local advances during the Vashon represented by three tills. Also included in the Vashon are glacio-fluvial and ice-contact deposits. Till in the study area is probably Vashon and not the later Sumas (J.Clague, pers. Comm., 1978). Mathewes (1973) dated post-glacial marine sediment in the study area at 12,690 Y.B.P. This date supports the conclusion that the later Sumas drift was not deposited as high nor as far to the west, as the study area. Armstrong (1957) mapped most of this area as "pre-Tertiary bedrock at or within 10 ft (3 m) of the surface, commonly overlain by till or outwash."

Particle size analysis of the Vashon till by Armstrong (1957) showed fractions: 57% sand, 41% silt, and 2% clay (USDA standards: clay, less than 0.002 mm; silt, 0.002 to 0.05 mm; and sand 0.05 to 2 mm). These analyses were done on lowland samples which Armstrong noted as being less sandy than those of similar age in mountain valleys. Distributions of this order were supported by deVries and Chow (1973, 1978) who found a particle distribution of 67.6% sand, 26.5% silt, and 5.8% clay for till in the Seymour Watershed. Thus, previous work implies that tills in the study area are generally coarse grained and probably over 60% sand.

Armstrong also noted that most of the clay sized particles consisted of fragments of quartz, feldspar, and other rock forming minerals. The high sand content and the lack of platey clay minerals probably create hydraulic conductivities in the mountain valley Vashon till that are greater than those expected for till in general.

Klinka (1976) mapped the soils of the UBC Research Forest. He found that the predominant soil class was humic-ferric podzol. Texturally, most soils were quite coarse with Sandy Loam being typical.

3.2 The Research Plot

3.2.1 Location and Physiography

The research plot is a 30 x 30 m hillslope stream-system located near Loon Lake (Figure 3-1) at an elevation of 354 m. It has a west-southwest aspect and an average slope of 22° which varies within the plot from almost flat to over 40°. Vegetation is very dense consisting of Douglas-fir and Western Hemlock 5 to 7 m high with dense underbrush of fern, salal, blueberry, etc. The area was clearcut in 1958 and has remained unthinned since then.

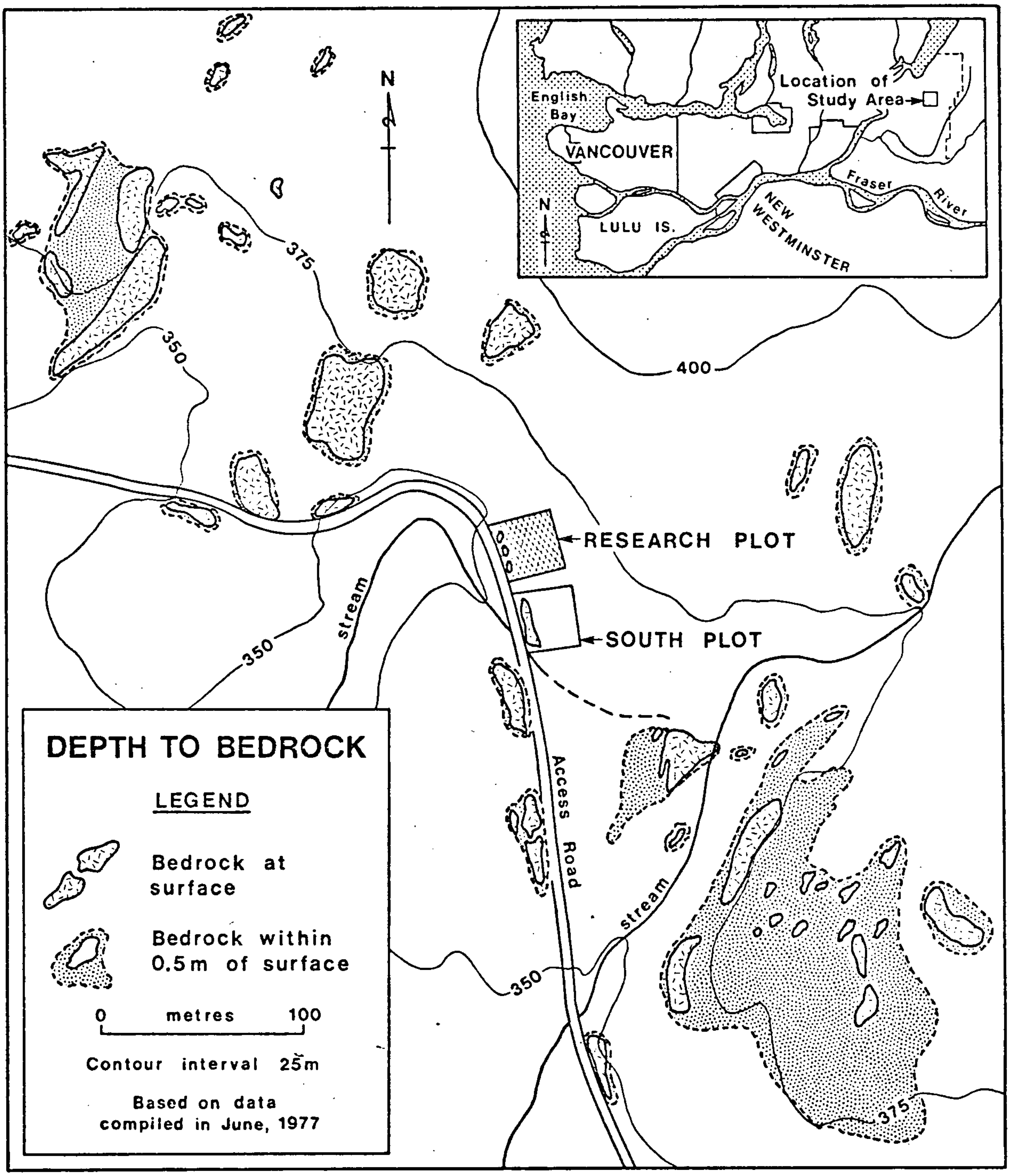

FIGURE 3-1 Depth to Bedrock and Location of the Research Area

3.2.2 Plot Pedology

The soil in the research plot has been analyzed by Nagpal and deVries (1976) and Bryck (1977). They described it as a humic-ferric podzol, sandy loam in texture. This is consistent with mapping done by Klinka (1976). A diagrammatic cross-section is presented in Figure 3-2.

The top layer of the soil is the organic forest floor. It is 0.05 to 0.25 m thick and is comprised of leaves, branches, roots, etc. in various states of decay. Some of this is slash left over from logging with branches and stumps up to 0.5 m in diameter. This unit is extremely permeable.

Underneath the forest floor is a discontinuous Ae horizon, 5 to 10 mm thick. This is underlain by 0.3 to 0.8 m of red-brown B horizon. This unit is very heterogenous with particles ranging in size from clays to boulders 1 m across. The matrix (particles under 2 mm) is texturally a sandy loam. Also contributing to heterogeneity are many roots and root channels. These are found in high concentration throughout the unit with a very dense mat of roots sometimes found along the surface of the underlying till. This occurs where the contact is well defined.

In these places the till and the B horizon are easily distinguished by differences in colour and hardness. Where these differences are well defined, a large contrast in permeabilities has caused downward growing roots to fan out along the contact, creating a root mat 2 to 3 cm thick. This phenomenon is self perpetuating as roots growing into this region raise the hydraulic conductivity, allowing for easier water extraction and therefore a better environment for more roots to concentrate.

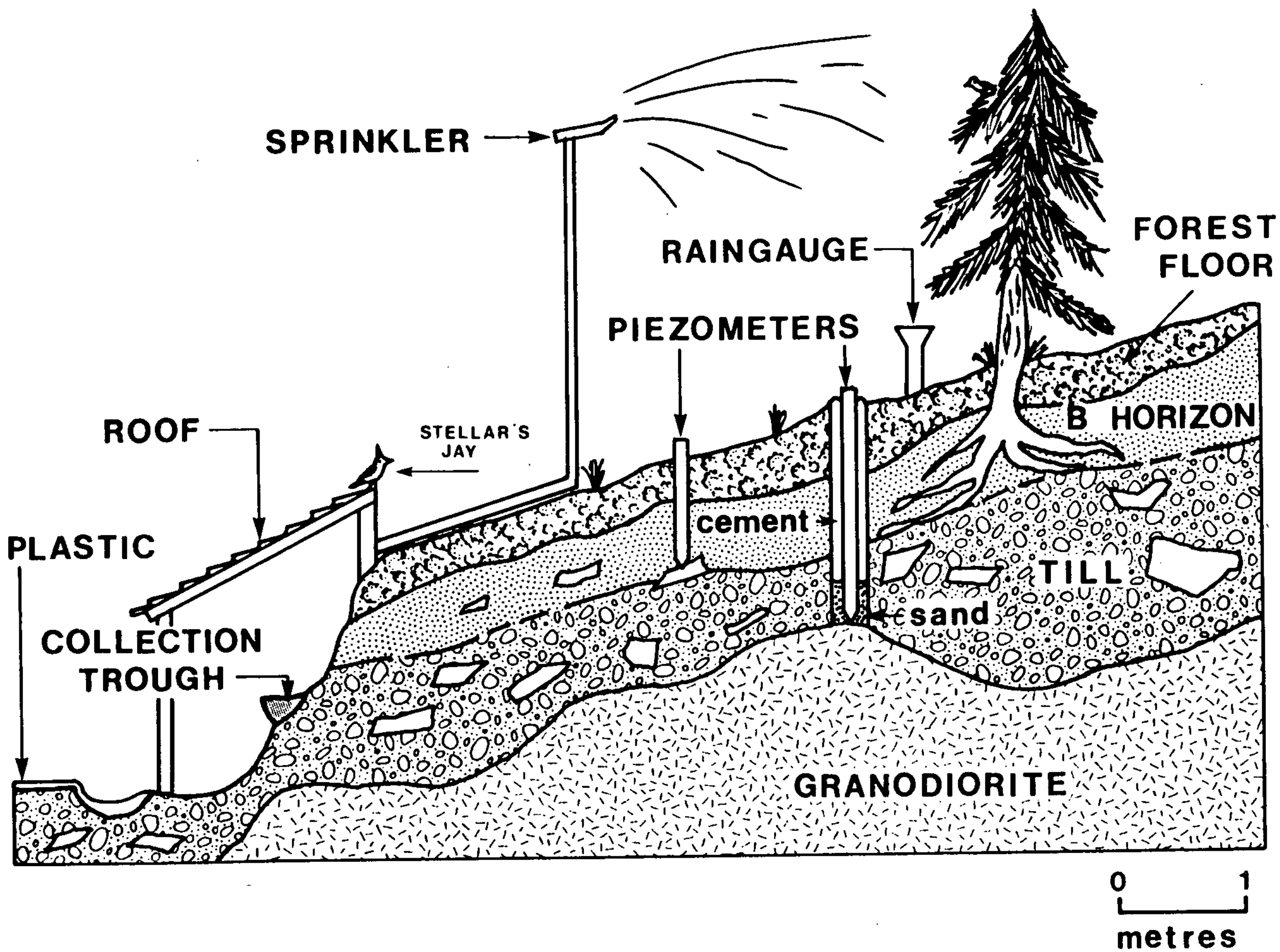

FIGURE 3-2 Diagrammatic Cross-section of the Research Plot

This situation results in two distinct hydrologic units: the B horizon with high conductivity because of the large concentration of organic material and the low-conductivity till. Where this contrast occurs, a perched water table often exists as was seen in several locations in and near the research plot.

In other places, a sharp contact does not exist. Rather, there is a gradual change from soft to hard and from red-brown to grey over a distance of one meter or more. Where this occurs, the till and the B horizon form a single hydrologic unit which has conductivities that increase toward the surface due to increasing concentration of organic pathways and decreasing degree of consolidation. It was observed that this situation predominates in the study plot and is therefore of major hydrologic significance.

3.2.3 Plot Geology: Till

As the nature of the contacts would indicate, the till itself is variable. It grades from grey to green-grey in an unweathered state to red-brown where extremely weathered. Where a root has penetrated its surface, a red-brown weathered zone a few centimeters wide and up to 0.5 m in length is sometimes surrounded by fresh, unweathered till. Some parts of the till are reasonably soft while other parts are so hard they can barely be broken with a pick. This variability is discussed more fully in a following section.

The till is usually found at a depth of 1 m. This depth is variable, in part from the original post-glacial and current topography and in part from the uncertainty of defining the contact. Exposures along the seepage face and isolation troughs around the plot show this variability in depth. In order to expand these two-dimensional sections to a three-dimensional surface, an attempt was made to probe the soil with a steel rod. This rod was hammered into the ground with the hope that differences in hardness would indicate a contact. Unfortunately, this technique was not successful as a marked contrast between the till and the B horizon did not always exist. The lack of a well defined contact also precluded the use of geophysical methods to locate the contact.

The till, like its daughter the B horizon, is poorly sorted. It contains particles from clay size up to angular boulders 1 m across. To quantify the particle size distribution, seven samples were analysed. Three of these came from the research plot while the other four came from locations up to 5 km away. All samples were air dried and gently crushed to break up aggregates. Non-matrix material was removed following which samples were Rotap sieved into 7 size fractions: 4.76, 2.00, 0.421, 0.210, 0.106, 0.074, and smaller than 0.074 mm. The under 0.074 mm fractions were then separated using the hydrometer method (Day 1965). A total of 18 size fractions were separated from each sample.

The average for these samples: 83.9% sand, 8.4% silt, and 7.7% clay (USDA standards), is considerably sandier than the 57% sand noted by Armstrong (1955, 1957) for lowland Vashon till.

However, it is consistant with his statement that mountain valley tills contain more sand than their lowland counterparts. It is possible that uncrushed silt-clay aggregates make up part of the sand size fraction, even though every attempt was made to minimize this possibility. Consistent results between duplicated samples tends to indicate the validity of the technique. A lack of variation between the seven locations indicates that the till is laterally homogeneous in particle size distribution.

3.2.4 Plot Geology: Bedrock

Directly underlying the Vashon till at variable depth is the bedrock granodiorite of the Coast Crystaline complex. This rock is hard, fresh and relatively unweathered. In the research plot, it is found at a depth of more than 1 m but usually not more than 3 m below the undisturbed surface. Along the man-made seepage face of the research plot, a cross-section of the till-bedrock contact can be seen. It varies from a depth of 1 meter in the northern half to below road level in the southern half. This reflects a depth of more than 3 m from the undisturbed surface. Isolation troughs and soil sample holes within the plot also indicate that depth to bedrock is usually around 2 to 3 m. The rod-hammer probe discussed above was also inconclusive for locating the three dimensional till-bedrock contact as large boulders in both the B horizon and till were indistinguishable from bedrock. Thus, the till-bedrock interface is not accurately known.

Where bedrock outcrops along the seepage face, fractures are widely spaced. This is the only place where bedrock is exposed in the plot and would, if representative, imply that bedrock hydraulic conductivities are low. However, the bedrock in the paired research plot to the south is well fractured along its surface and must have relatively high conductivity as previous work by deVries (pers. comm. 1977) indicated that very little applied irrigation water appears as output. Thus, the hydrologic role of bedrock in the current research plot could not be surmized from geologic investigation as there was no way to tell fracture density beneath the surface. As previous experimental work (Nagpal and deVries, 1976) indicated a high leakage rate, it was at first assumed that fracturing was significant. Further work reported in a later section does not support this conclusion.

3.3 Near Plot Geology

Because this report is process oriented, it was felt that the generality of the research plot should be established. Two techniques were used to investigate the hydrogeological conditions of the research plot vis-a-vis those of the area: surficial mapping and an examination of a series of vertical road cut sections.

3.3.1 Surficial Mapping

An attempt at surficial mapping was made in the area around the research plot. This was done using a rod-hammer probe, an Oakfeld auger, and a shovel. Because the soil contained many large rocks and was often covered by 1-2 m of slash, these techniques were seldom useful. It was possible, however, to map the regions with bedrock at or near surface (within 1/2 m) and those regions with bedrock at greater depth. It was postulated that in areas with deeper bedrock, hydrogeologic conditions would be similar to those of the study plot. In shallow bedrock areas, it is possible that flow mechanisms are different. The results are shown in Figure 3.1. It can be seen that in 60 to 70% of the area around the study plot, depths to bedrock are similar to those within the plot. This technique is limited in that it does not indicate the actual depth or nature of bedrock-till or till- B horizon contacts. To augment this, a vertical section technique was used.

3.3.2 Vertical Profiles

Since the mapping technique used did not show actual thickness-contact relationships, 23 vertical road cut sections within 2 km of the study plot were examined in detail. In addition, two other sections from the Seymour Watershed were reconstructured from the literature (Chamberlin, 1972; deVries and Chow, 1973, 1978). The depths and thicknesses of the forest floor, B horizon, till, and bedrock were noted. These are presented in Figure 3-3. Sections one through 23 represent

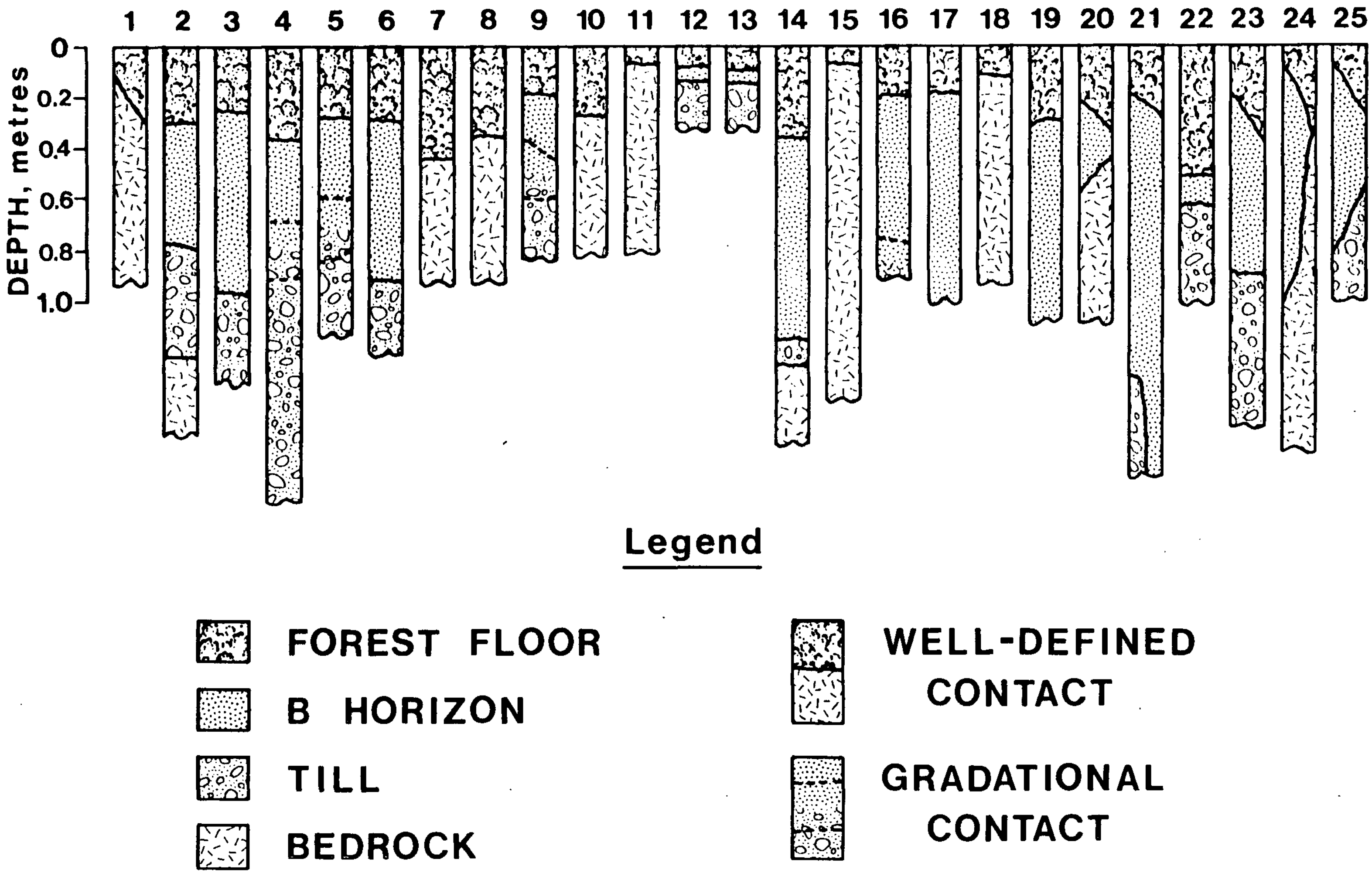

FIGURE 3-3 Twenty Five Vertical Profiles from near the Research Plot

profiles near the research plot, while section 24 is based on Chamberlain (1972) and section 25 is based on deVries and Chow (1973). It can be seen that 12 out of 25 sections had definite to semi-definite (± 0. 15 m) contacts, six had broader gradational contacts (usually over 0.50 m wide) and seven sections had forest floor resting on bedrock, or at most with a very thin B horizon interspaced between forest floor and bedrock. Thus, it can be seen that relative thickness of the four units in the research plot are fairly typical of those in the area.

The forest floor material was similar to that seen in the research plot. It contained decayed and semi-decayed twigs, branches, leaves, etc., 0.1 to 0.45 m thick. In all cases, it appeared to be very open and permeable.

The B horizon was also similar to that seen in the research plot. It was brown-grey to red-brown in colour and varied from 0.05 to greater than 1 m thick. It had many live roots as well as decayed root zones. As these were areas of higher water flow, greater weathering was often seen as indicated by red-brown limonite-hematite stains. The presence of these weathered zones near the bottom of the soil profile made identification of the contact with the till difficult, as these zones often fingered into what was unquestionably till.

The B horizon was usually not saturated. This was not surprising as the study was conducted during June following a record dry winter. However, in 4 of the 25 locations the B horizon was saturated and had observable discharge. In two of these saturated areas, a perched water table existed above 0.3

to 0.5 m of unsaturated till. This situation was also observed at one location in the research plot.

The till in the surrounding area was similar to that in the research plot. It ranged in thickness from as little as 0.1 m to more than 0.60 m. It may have been thicker but there were only two locations where it was possible to see both top and bottom. Its colour ranged from an unweathered grey to a weathered red-brown. It was also seen to be very hard in some places and soft in others. This difference in hardness was also observed in the study plot.

Texturally, the nearby till was similar to that in the research plot but with one exception. In several locations a thin (1-3 cm) layer of clay was present at the till-bedrock contact. This layer, which undoubtably reflects some ice-rock interaction, may be significant hydrologically in that the clay would form a local low-conductivity zone and tend to minimize flow into bedrock joints. This clay layer was not observed on any bedrock with joints, however.

In short, the till in the research plot is similar to that in the surrounding area, especially in the variations exhibited. This variability, caused by differences in degree of weathering, led previous workers to the conclusion that both an ablation and a compacted till exist (Nagpal and deVries, 1976; Bryck, 1977). However, this is probably not the case. Nowhere could two distinct tills in contact with each other be seen. Nor did there seem to be any differences in pebble fabric between the softer and harder tills. Detailed examination of the 23 vertical sections showed that zones where the term ablation till could be applied graded into compacted zones. Particle size analysis of both hard and soft tills yielded the same results. It may be possible that two different tills did exist but, as colluvial processes have been very active, any differences which were once present are no longer seen. Thus, it is concluded that only one till is present in the study area.

Hydrologically, the question of whether there are one or two tills is unimportant. Hydrologic behavior is not based on geological history, but on actual physical characteristics. Soft, weathered till with organic channels behaves like B horizon and can be included with it in one hydrologic unit. Thus, a separation of hydrologic units is based on organic content and degree of compaction, and not on genesis.

The bedrock in the vertical sections was all granodiorite. Depths ranged from 0 to more than 2 m below the undisturbed ground level. In some locations, it was not possible to determine the total depth as the bedrock contact was below the surface of the road cut. In depth to bedrock, Figure 3-3 shows that the study plot is representative for 18 out of 25 sections.

Most bedrock exposures were more fractured than those seen in the research plot. In several places large surface fractures 2 to 3 cm across could be seen (Figure 3-4). Large fracture apertures combined with fracture spacings of 10/m could give a relatively high hydraulic conductivity. The origin of these fractures is not known. Their orientation appeared to be random and as the only places where well-exposed rock faces could be viewed were road cuts, it is possible that these fractures were caused by road building. Origin aside, it is probable that

FIGURE 3-4 Bedrock Fractures
Tape length: 0.37 m

fracture apertures become much smaller with depth. This decrease in size coupled with the clay rich till along the contact, suggests that flow into and through the bedrock should be minimal and most likely confined to shallow depths. Thus, the differences in fracturing seen in the plot and surrounding areas do not limit generalization of the experiment to the study area.

4.0 EXPERIMENTAL PROCEDURE

In order to examine the mechanism of stormflow generation, a controlled irrigation experiment was conducted. This experiment consisted of monitoring water inflow, water outflow, and piezometric pressures in a study watershed during rising, steady-state and falling outflow conditions. The experiment was conducted on the 8th through 23rd of August 1977. The watershed consisted of a 30 m square section of a hillslope, an artificial stream, rain gauges, piezometers, and outflow tipping buckets. In addition, chemical tracers were introduced during the first part of irrigation with samples taken at four stages of the system: at the source, on the hillslope, at the "basin" outlet, and below the hillslope. These samples were examined in order to deduce flow paths taken by rainfall on its journey to the stream. In this chapter, the instrumentation and procedure of the experiment are discussed. In addition, the auxillary measurements that were made to calculate hydraulic conductivities are described. A diagrammatic representation of the hillslope, stream, and instrumentation used in the experiment, is included in Figure 3-2.

4.1 Modifications to the Plot

In order to increase the signifigance of the experimental results, several modifications to the natural site have been made. A roof over the artificial stream was made to eliminate direct precipitation into the "stream" and the subsequent measurements of water not flowing through the hillslope .

Another modification was an isolation trench around the sides and back of the plot, dug down to the level of the till. The trench was excavated in order to reduce the unmeasurable addition of subsurface flow from outside the plot. Unfortunately, because the soil was so stony, the trench could not be dug deep enough. A final modification was the construction of a concrete runoff trough and placement of a 4 mil plastic sheet below the plot to drain off sprinkler overspray. These modifications ensured that below-plot piezometer rises were due to subsurface flow originating from irrigation water applied to the study plot.

4.2 The Irrigation System

Water was applied to the research plot via an eight sprinkler irrigation system. Placement and sprinkler design was such that "rainfall" was as uniform as possible. The sprinklers rose approximately 6 m above the ground surface and, for all but a few of the tallest trees, were 1/2 meter or more above crown height. Water was pumped from an aerated sewage treatment lagoon 500 m west of the site.

Using a metered pump, it was planned to irrigate at a constant rate throughout the experiment . However, plugging of the intake screen by algae sometime before the morning of the third day of irrigation (August 10) caused a "rainfall" rate which decreased continually until 2:30 hrs on the fifth day (August 12), when the situation was discovered and corrected. Luckily, this problem turned out to be of benefit to the experiment because it produced a hydraulic wave which was measurable throughout the system.

The location of the sprinklers is indicated in Figure 4-1 by the letter s.

4.3 Rain Gauges

Ten collection gauges and two tipping bucket continuous-recording rain gauges were used to monitor input to the study plot. In order to measure natural rainfall which could have occurred in addition to irrigation , one continuous recording rain gauge was monitored outside of the irrigated area.

4.3.1 Collection Type Rain Gauges

Collection type rain gauges were placed 1/2 m above the ground surface at 10 locations within the study plot. These consisted of beveled edge plastic funnels with collection areas of 82.52 cm^2, feeding sealed storage bottles. Water in these bottles was measured once or twice daily with a one liter graduated cylinder.

Rainfall rates at the ground surface varied according to position in relation to the sprinklers and amount of vegetation cover. In order to calculate average rainfall as accurately as possible for the entire plot, Thiessen weighted polygons were used (Dunne, 1974). Rain gauges were placed such that areas defined by Theissen polygons were approximately the same as areas defined by vegetation cover. Thus, each rain gauge was representative of precipitation that actually hit the surface in its respective area. From the collection gauges, average

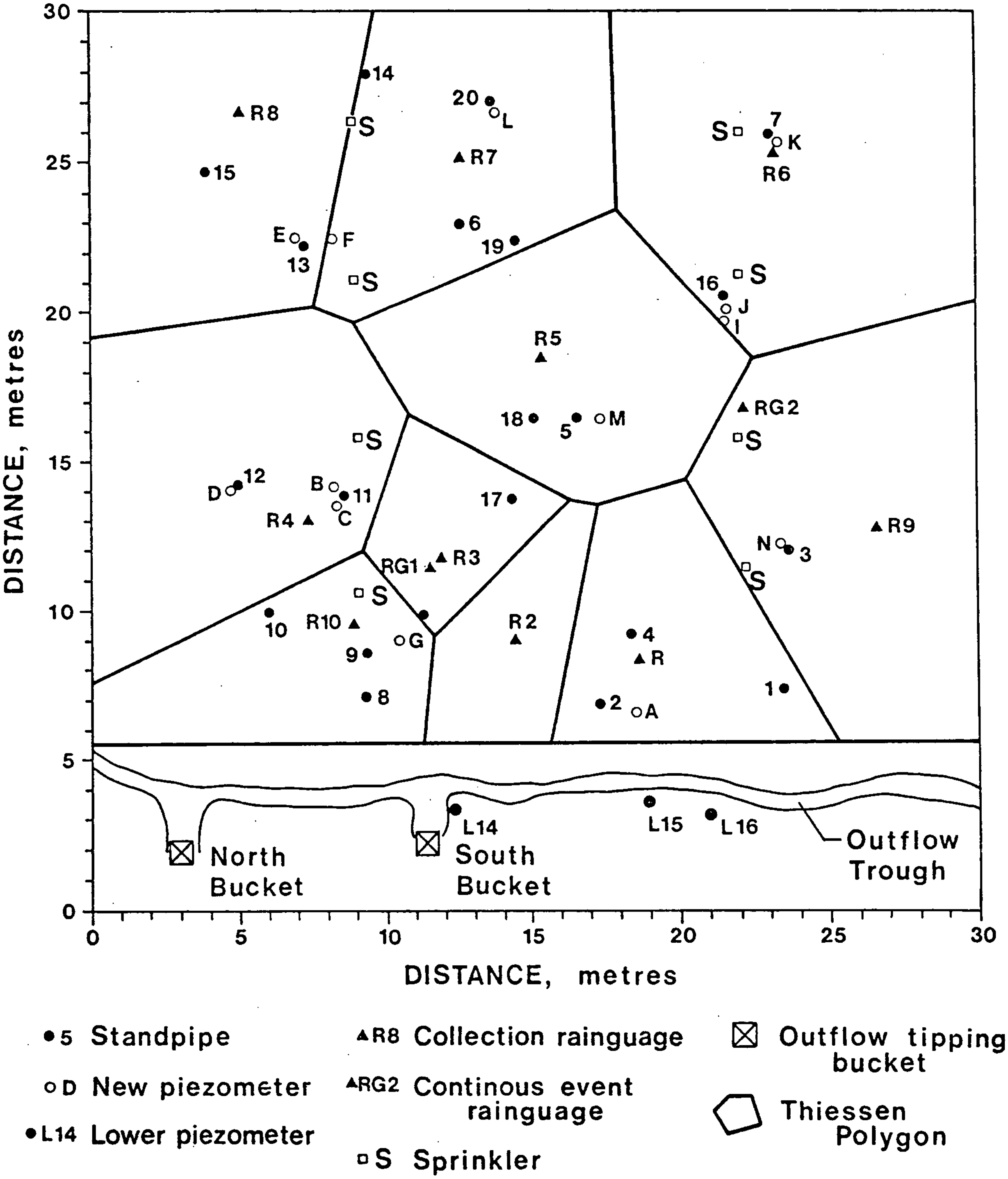

FIGURE 4-1 Layout of the Research Plot

rainfall rates for the entire plot were calculated. These rain gauges, labeled R1 to R10, and Thiessen polygons are shown in Figure 4-1.

4.3.2 Tipping Bucket Rain Gauges

In order to corroborate collection rain gauges and to monitor diurnal variation, two continuous tipping bucket rain gauges were used. These had a diameter of 25.4 cm and an area of 506.7 cm^2. Monitoring was done with an Esterline-Angus, paper-roll, continuous-event recorder with each event equal to 16.0 cm^3 (0.0316 cm rain depth equivalent).

Mean rainfall rates were calculated by counting the number of events and estimating to 10% the fraction of the uncompleted event in each hourly period. Resolution was ± 4% for low rainfall rates and as fine as ± 1/2% for the highest rainfall rates. This degree of resolution made it possible to detect and measure diurnal rainfall variation. Data from these rain gauges were not used directly to calculate input rates. Rather, the measured percentage diurnal variation was superimposed on the values obtained from collection type gauges. Thus, it was assumed that diurnal variation was uniform throughout the plot.

Figure 4-1 shows the location of the continuous event gauges, labeled RG1 and RG2.

4.4 Piezometers

In order to understand hillslope flow paths, 50 piezometers were used. Twenty of these had been installed and used by Nagpal and deVries (1976) for earlier infiltration studies. For convenience these are called standpipes. An additional 30 piezometers were designed, built and installed by the author in order to help examine flow in the till and flow out of the research plot . These are called the new piezometers .

4.4.1 Standpipes

Twenty piezometers were installed in the B horizon by Nagpal and deVries to study the water table configuration . They were made of 1.91 cm O.D. galvanized steel pipe approximately 1.45 m long and had ten 2 mm holes in the lower 1/4 m for an intake screen. These were not true piezometers as they were not sealed at the tip and therefore did not measure head at a point. However, they were sealed at the surface with clay to prevent "stem" flow down the tube. These piezometers , or standpipes as they are more accurately termed, measured the water table level in the permeable B horizon .

Their locations are shown in Figure 4-1 numbered 1-20.

4.4.2 New Piezometers

The 30 piezometers installed by the author consisted of 4.45 cm O.D. galvanized steel conduit, 1.45 to 4.01 m long. They were tapered at the tip to prevent clogging during installation and had 34, 1 mm by 30 mm slits in the lower 13 cm for the intake screen (Figure 4-2).

A two-person Groundhog power auger was used to drill the 6 cm hole required. Because the terrain was both steep and rugged and especially because there were many cobbles and boulders in the B horizon and till, it was not possible to drill deeply enough to locate these piezometers where they could unequivocally establish flow rates in the till. (For this reason I do not recommend this method of installation for those considering research in similar locations.)

After drillng, sand was placed in the bottom of the hole to ensure hydrologic coupling. A volume of sand calculated to fill to a depth just above the intake screen was applied with a tube inserted to the bottom of the hole. The piezometer was then driven into the sand and sealed into place.

The sealing of these piezometers was attempted in two ways. The first nine were sealed with bentonite expanding clay (Quickgel). This was attempted with both dry and slurry forms. It was not certain with either of these methods that sealing took place just above the intake screen. For this reason, some of these piezometers may, in effect, be standpipes. Fortunately, only one of the questionable piezometers (A) was in the research plot . Five of the remaining eight were so far below the plot, they did not respond to irrigation and so sealing was unimportant. The remaining piezometers were well-sealed with a concrete slurry.

Fourteen piezometers were intstalled in the research plot to better understand the role of the Vashon till. These were placed in nests of two or three which usually included one of the

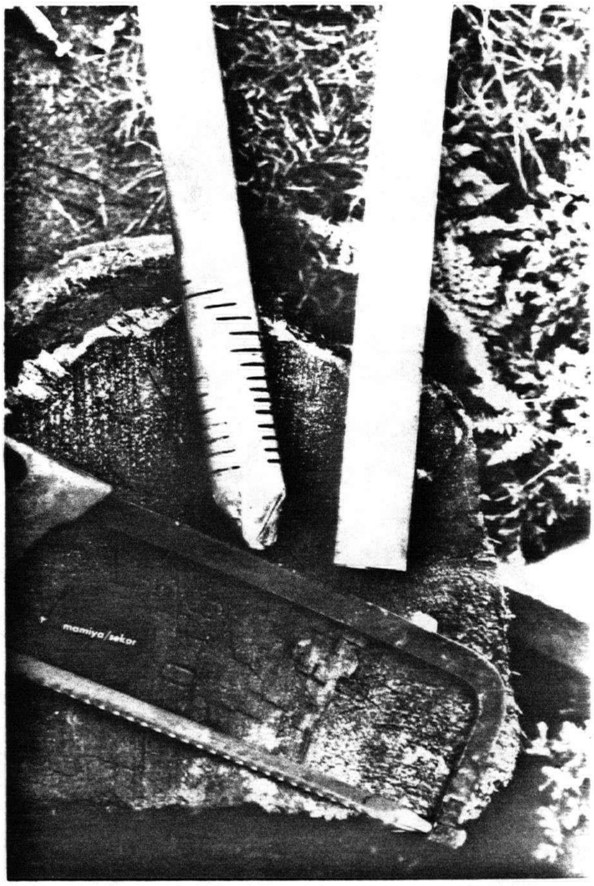

FIGURE 4-2 The New Piezometer

standpipes. From these nests, vertical gradients and response times as functions of depth were examined. Most of these piezometers were placed within the till, but unfortunately none greater than 0.5 m below the B horizon-till content. The locations of the new in-plot piezometers are indicated in Figure 4-1, labeled A through N.

To study leakage from the irrigated area, 16 piezometers were installed below the study plot: three just below the collection trough, eight below the road near the natural stream, downslope from the plot and the last 5 at some distance (up to 125 m) away. Because the lower five showed no response, they are not discussed in this report. The other 11 piezometers are labeled L6 through L16 in Figure 4-3.

4.4.3 Reading the Piezometers

Piezometer water levels were measured using an acrylic platic tube containing an expanded polystyrene (Styrofoam) float which adhered to the tube wall via surface tension at the height it was floating. The measuring tube was inserted into a piezometer and then withdrawn with the distance from the tip to the float measured to the nearest millimeter. Because soil in the water caused the position of the float to vary, readings were taken two to five times for each piezometer to establish a representative water level. Accuracy of the average obtained is estimated to be ± 0.5 cm.

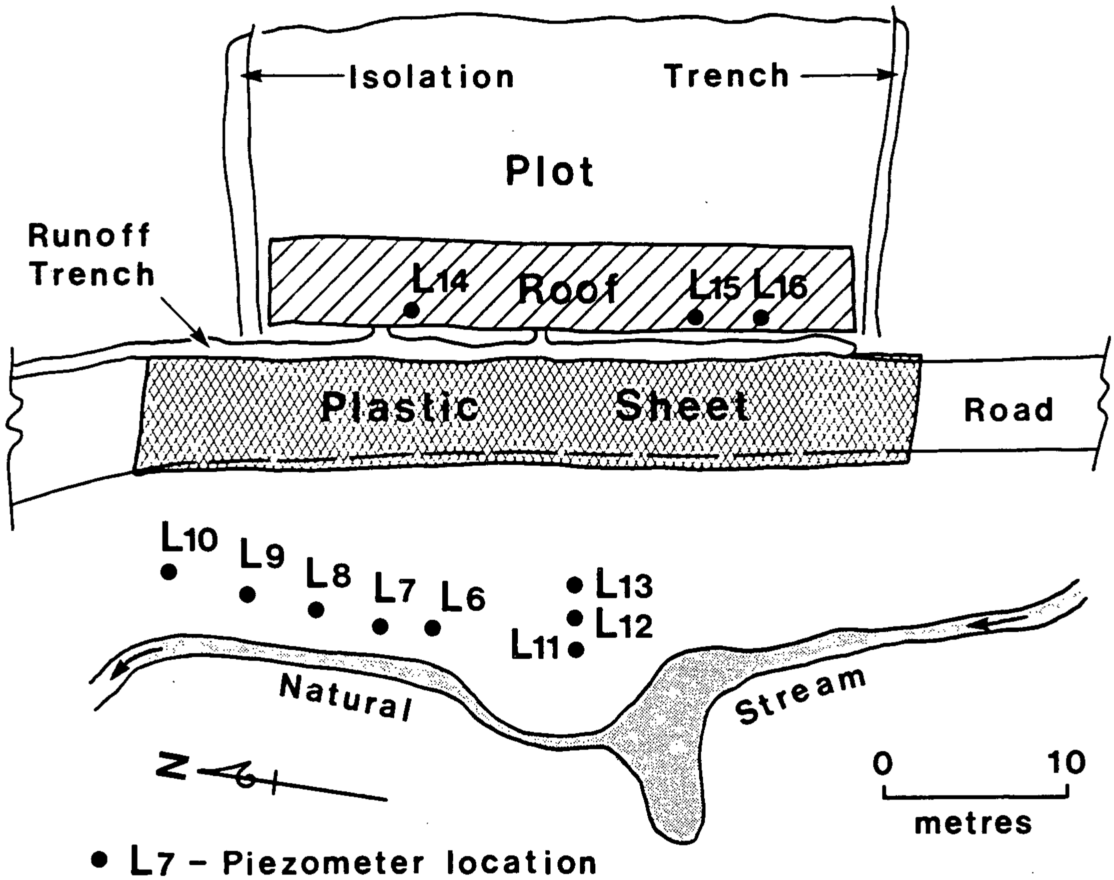

FIGURE 4-3 Location of the Piezometers Below the Research Plot

Readings were taken at 90 minute to half day intervals, depending on whether the experiment was in a transient or steady-state condition.

4.5 Collection Trough and Tipping Buckets

Outflow from the seepage face (Figure 3-2) was collected by an artificial stream (concrete trough) on the surface of the till, 2 m below the original ground surface. Although this trough had been constructed for previous experiments by Nagpal and deVries (1976), it was overhauled to reduce the possibility of loss either at the trough-till interface or through cracks in the original concrete. This trough was constructed with two outlets such that outflow from the southern 2/3 of the study plot was measured separately from the northern 1/3.

Outflow was measured with calibrated tipping buckets. These were constructed with adjustable volumes of up to 2.3 litre per tip. The north bucket was calibrated at 1.96 litre per tip while the south bucket was originally set at 2.25 litre per tip. This volume did not remain constant however, as the adjustment screw shifted twice during the experiment. To compensate, a correction factor was used for the periods during which no actual calibration existed. This factor was based on the observation that during both calibrated periods, the average outflow measured by the south bucket was approximately 1.5 (± 0.3) times the outflow measured by the north bucket. Maintaining this ratio, five calibration volumes for the south bucket were calculated. With these, the actual timings and relative volumes (within each calibration period) were preserved while a fairly accurate estimation of outflow volume was made possible. The calibration of the north bucket remained constant throughout the experiment.

Outflow was recorded on two channels of the same event recorder used for rainfall. Hourly outflow rates were calculated by counting the number of tips during the 15 minute period past each hour and multiplied by four. Because the number of events during this period was large (usually 80 to 140), it was neither necessary nor practical to estimate the remaining volume of water for the uncompleted tip at the end of each period. This led to a resolution of 1% to 4%, depending on outflow volume. Such precision is comparable to other elements of the system.

4.6 Chemical Tracers

In order to deduce flow paths through the hillslope, chemical tracers of known concentration were applied to the hillslope. These tracers consisted of Cl^-, K^+, NO_3^-, NH_4^+, and PO_4^{3-}. Tagged water mixed in a sewage treatment lagoon 500 m to the west was applied for the first seven days of the experiment, at which time a water-flux steady state was well established. Clean lake water was then applied for the remaining five days with hopes that the clean water "front" would be observable.

Samples were taken four times daily at four parts of the system: at the intake to the eight sprinklers, in the plot at piezometer C and B, at both outflow tipping buckets, and below the plot at piezometers L15 and L9. Electrical conductivity measurements were also made on the outflow water to observe solute breakthrough.

Samples were analyzed by the Pollution Control Engineering Laboratory, U.B.C., using an a autoanalyzer. Analyses were done within 48 hours, after being treated with sulfuric acid and stored near 0°C to reduce ionic species transformation. Electrical conductivity measurements were made in the field using a Yellowfield model 33 conductivity meter.

4.7 Hydraulic Conductivity Determinations

The hydraulic conductivity of the till and the lower B horizon was determined by several methods. Direct measurements were made by slug tests and infiltrometer tests. Values were also calculated indirectly from the results of the irrigation experiment .

Interpretation of slug test data is based on Hvorslev's (1951) method. This approach combines an empirical shape factor and the differential form of Darcy's law to produce formulae for conductivity as a function of head change over time from an injected slug of water. This test was performed on six of the new piezometers during a period of steady-state flow in the hillslope , well after the irrigation experiment was run. The homogeneous isotropic form of the Hvorslev equation was used. This approach should yield at least an order of magnitude accuracy. In addition to the piezometers tested, L15 was pumped for chemical analysis during the experiment . This allowed the use of the subsequent water level rise to be used in the Hvorslev equation as a bail test.

An infiltrometer, designed and built by J. DeVries and C. Paul with some modifications by the author, was used to measure infiltration rates on cleared-off sections of till and lower B horizon This device consisted of a 0.5 x 0.5 m grid of 400 hypodermic needles which produced artificial raindrops, supported approximately 0.8 m above a test section of till or soil. Water was supplied by a controlled burette system. The theory behind this instrument is based on Philip (1957) who stated that the steady-state infiltration rate is equal to the saturated hydraulic conductivity . Unfortunately, the infiltrability of the till was too low for the infiltrometer to maintain and measure the uniform just-ponding conditions required for the use of Philip's equation. To overcome this problem, higher input rates were used with an attempt to measure runoff and calculate infiltration rate from the difference between input and output. This technique did not yield satisfactory results as it was not logistically possible to run the apparatus long enough to ensure steady-state conditions . For these reasons, the results of this work are not presented in this report.

4.8 Data Reduction

Data analysis was simplified by the use of the computing facilities at the University of British Columbia . Raingauge data, read as a volume of water at an arbitrary time, were converted directly to average rainfall rates and infiltration volumes. Piezometers , which were read in several ways (i.e. relative to the top or to the bottom depending on depth to water) were converted to a common datum. Hydrographs for rainfall, piezometers , and outflow were drawn using <u>Calcomp</u> subroutines and printed by a <u>Tektronix</u> 4012 CRT display terminal with a hard copy attachment. Thus, many tedious calculations were avoided with the bonus of increased accuracy.

5.0 Experimental Results

In this chapter the results of the field rainfall-runoff experiment are presented. Hydrographs and tables for artificial rainfall (hereafter called rainfall), piezometric levels, and outflow are listed. These are followed by results from the slug tests and chemical analyses. Included with these data are minor discussion pertaining to basic interpretation . The major discussion of analyses of the field experiment are presented in the next chapter after all basic data have been presented.

5.1 Rainfall

Mean rainfall rates for each of the collection type rain gauges are listed in Table 5-1. Total plot input expressed both volumetrically and averaged over the irrigated area is shown in Table 5-2. Rainfall from continuous recording type gauges is graphed in Figures 5-1 and 5-2.

A diurnal rainfall variation was indicated by both recording gauges. In order to better demonstrate this variation, superimposed daily rainfall from RG1, for the period of August 13th through 20th, is shown in Figure 5-3. Minimum input occurred between 13:00 and 19:00 hrs. There was no single, well defined input maximum except on the 15th, 16th, and 17th when peak rainfall occurred at 10:00 hrs. The minimum coincided with the hottest part of the day when much mist and vapour was observed over the plot. It is likely that intense evaporation and low humidity around the plot, caused by daily temperatures near 30°C, produced the diurnal variation in rainfall reaching

Table 5-1. Average rainfall rates for the ten collection type rain gauges.

Rain gauge	1	2	3	4	5	6	7	8	9	10
August	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr	Cm/Hr
9.63	0.33	0.40	0.50	0.31	0.55	0.35	0.48	0.41	0.32	0.33
10.52	0.33	0.07	0.50	0.26	0.54	0.32	0.44	0.37	0.27	0.23
11.48	0.23	0.08	0.36	0.15	0.41	0.27	0.29	0.24	0.20	0.21
12.08	0.15	0.03	0.22	0.06	0.22	0.13	0.08	0.02	0.08	0.18
12.85	0.34	0.10	0.48	0.25	0.48	0.28	0.42	0.38	0.22	0.24
13.32	0.36	0.24	0.55	0.41	0.59	0.33	0.64	0.67	0.27	0.27
13.81	0.42	0.14	0.55	0.35	0.57	0.28	0.49	0.48	0.14	0.24
14.49	0.39	0.16	0.57	0.30	0.58	0.34	0.72	0.62	0.29	0.28
15.37	0.35	0.17	0.27	0.33	0.57	0.36	0.48	0.59	0.18	0.27
16.37	0.33	0.12	0.51	0.32	0.54	0.36	0.42	0.13	0.26	0.24
17.37	0.33	0.10	0.51	0.32	0.53	0.35	0.48	0.10	0.27	0.23
18.39	0.35	0.10	0.52	0.30	0.54	0.33	0.51	0.09	0.32	0.23
19.36	0.34	0.20	0.52	0.25	0.55	0.36	0.62	0.17	0.26	0.25
20.36	0.35	0.10	0.52	0.31	0.54	0.35	0.55	0.23	0.25	0.25
21.40	0.36	0.09	0.51	0.35	0.59	0.33	0.48	0.28	0.24	0.23
21.76	0.38	0.16	0.56	0.34	0.58	0.37	0.52	0.42	0.26	0.23

Table 5-2. Total plot input and average rainfall rate for the whole plot.

August	Total Input	Average Rate	
	m^3/hr	cm/hr	m/s(x 10^{-7})
9.633	2.386	0.33	8.2
10.524	2.053	0.28	7.8
11.480	1.462	0.20	5.6
12.078	0.617	0.08	0.2
12.851	1.956	0.27	7.5
13.323	2.757	0.38	10.6
13.813	2.253	0.31	8.6
14.493	2.691	0.37	10.3
15.372	2.292	0.31	10.6
16.369	1.906	0.26	7.2
17.368	1.902	0.26	7.2
18.394	1.967	0.27	7.5
19.360	2.065	0.28	7.8
20.365	2.076	0.28	7.8
21.396	2.128	0.29	8.1
21.764	2.340	0.32	8.9

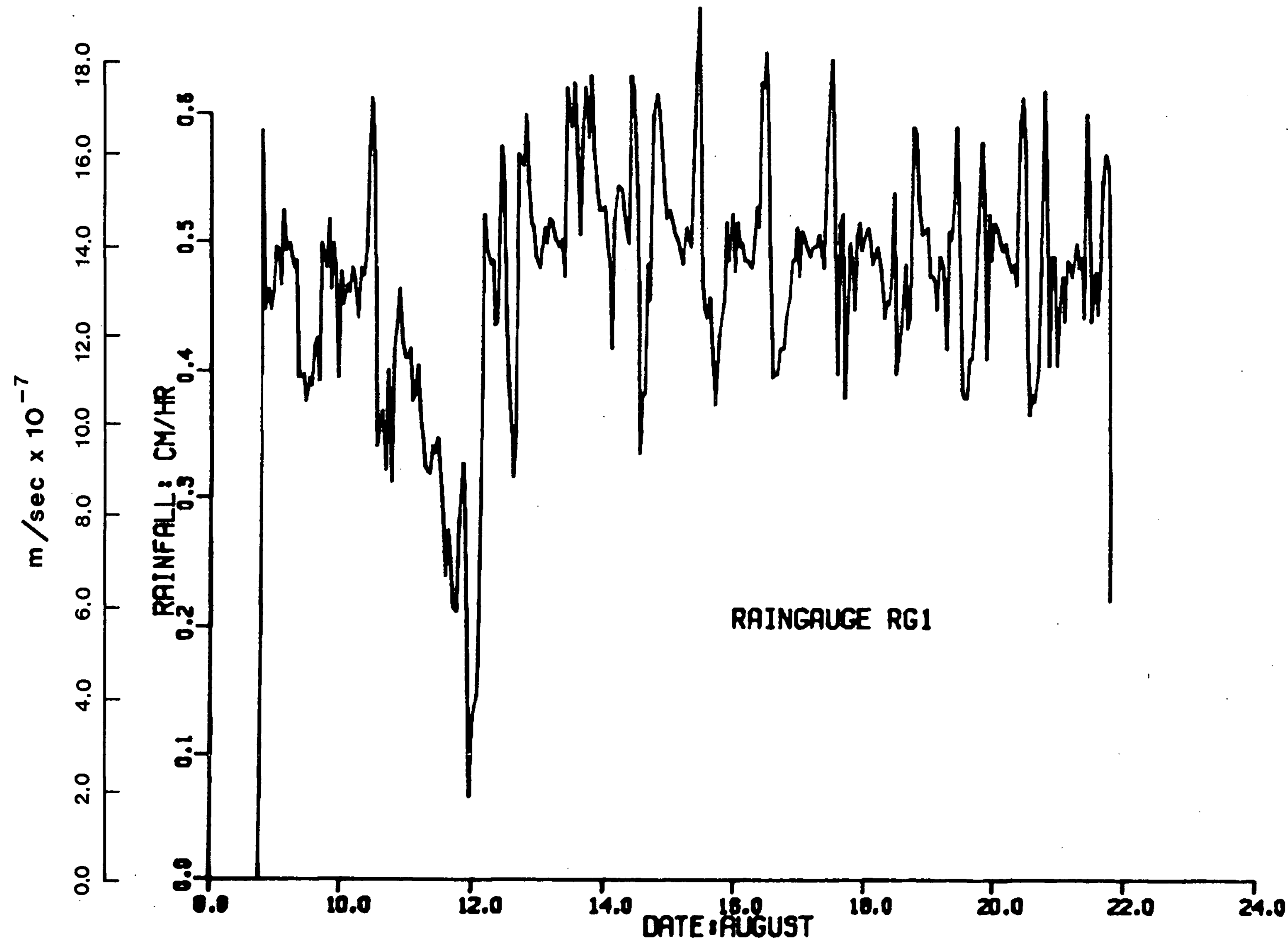

FIGURE 5-1 Rainfall in raingauge RG1

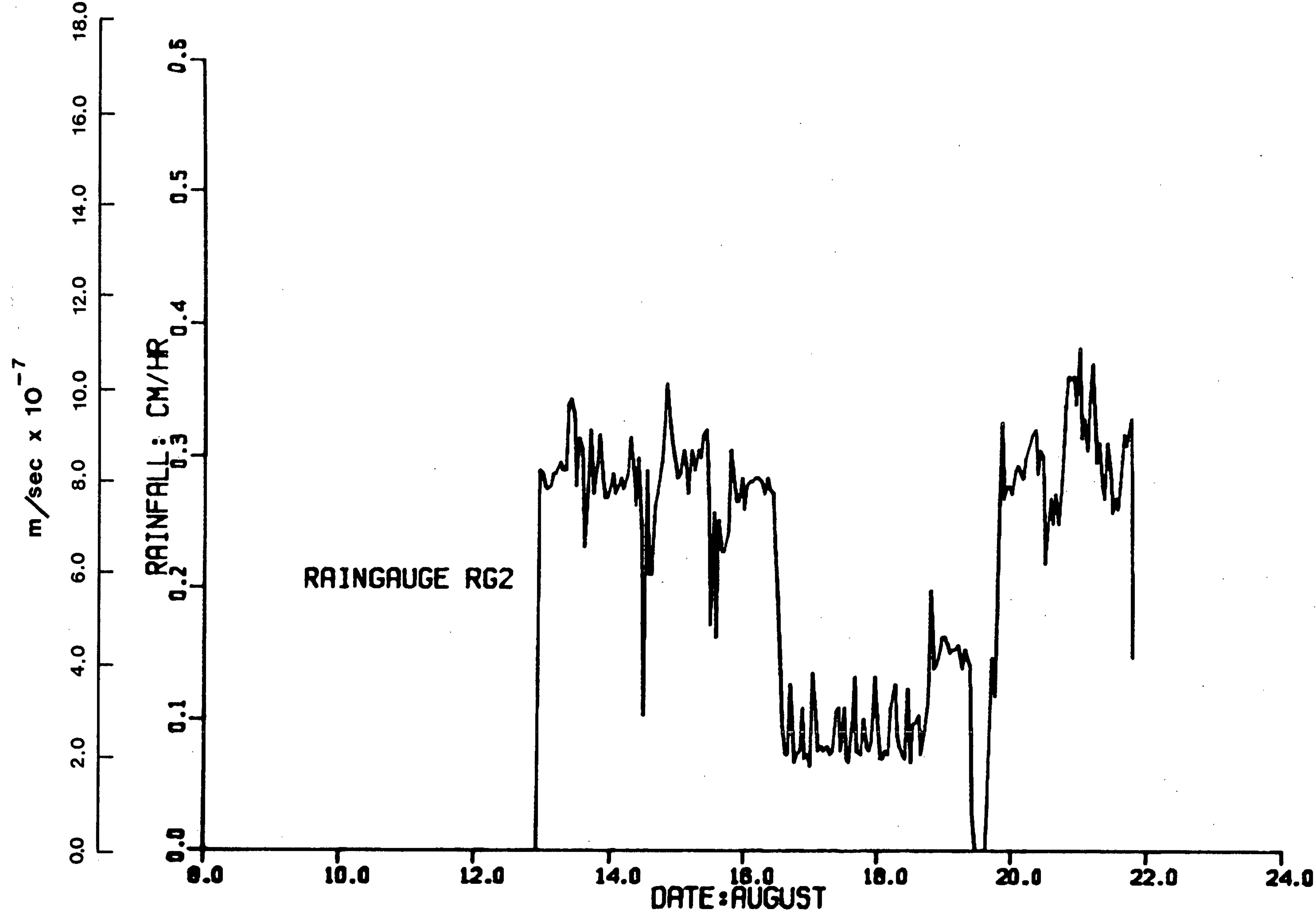

FIGURE 5-2 Rainfall in raingauge RG2

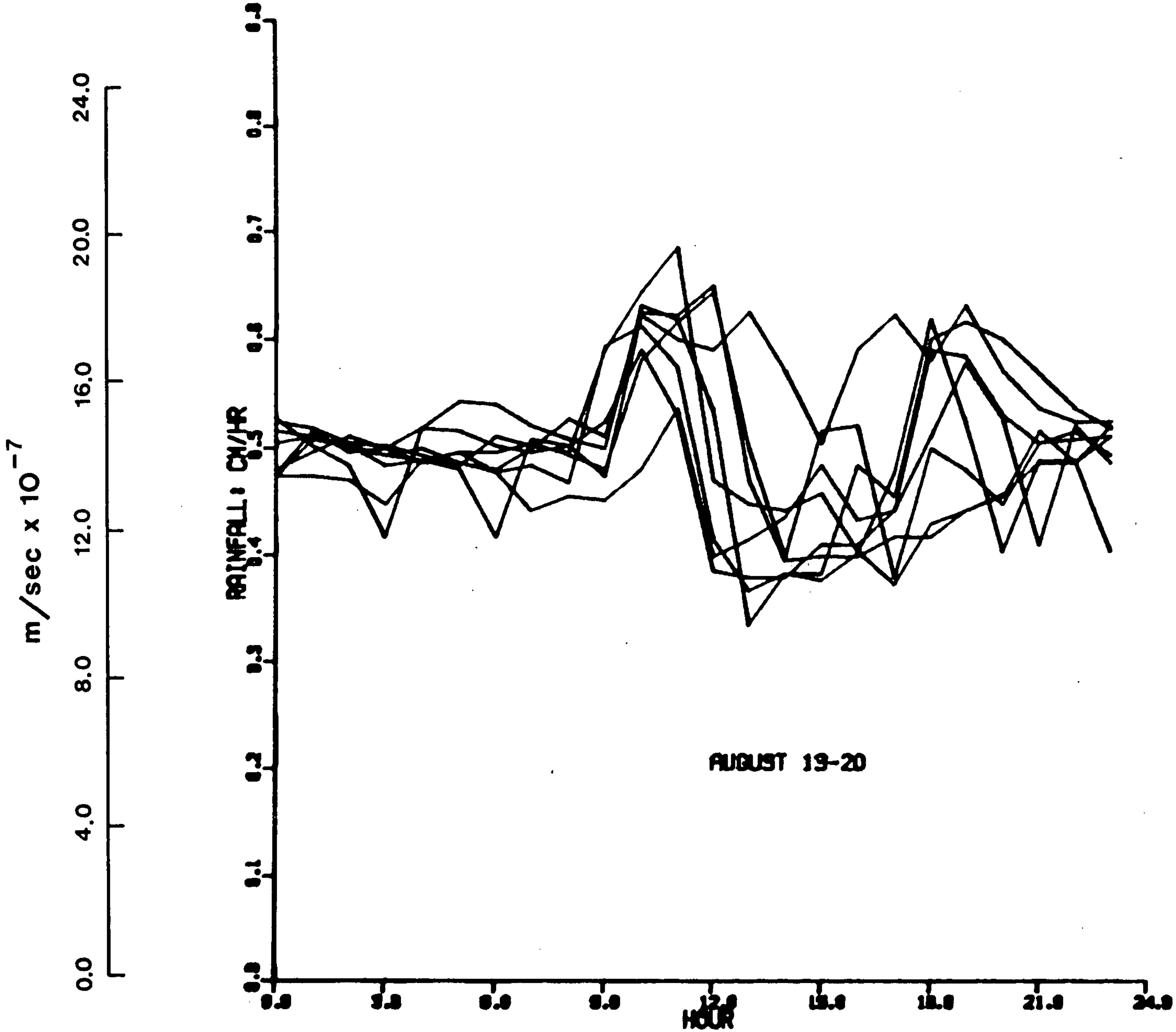

FIGURE 5-3 Superimposed daily rainfall from rain gauge RG1

the surface.

The collection and continuous rain gauges gave similar results. Both types of gauges indicated the decrease in precipitation because of the partial clogging of the irrigation pump input screen. Also, collection gauge R3 and recording gauge RG1, located near each other (Figure 4-1), both indicated a mean rainfall rate of 0.52 cm/hr. This aggreement indicates that calibration was consistent for both types of gauge. Seemingly contrary to this conclusion, RG2 showed a marked decrease in rainfall during the period of the 16th through the 20th that was not seen in surrounding collection type gauges. Such behavior was due to malfunctioning of the mercury switch on the tipping bucket. Therefore, this precipitation decrease was not real.

5.2 Piezometers

Piezometer hydrographs are shown in two ways: individually and in groups of two to five. These groups are not true vertical nests because the horizontal spacing between some of these piezometers is greater than their vertical separation. All piezometers were referenced to a common datum at the road (0.0 m). The piezometer hydrographs are listed in the Appendix, Figures A5-1 to A5-23.

Most piezometers within the plot (1 through 20, B through M excluding F and K) were in the B horizon and responded quickly to the initiation of rainfall. Piezometer 19 rose within 1 1/2 hours while two others (3 and 20) began to rise in less than 3 hours. The mean initial response time was 10.3 hours for the standpipes (shallower) and 16.9 hours for the new piezometers (deeper). The later response of deeper piezometers is contrary to unsaturated infiltration theory in homogeneous media (Rubin and Steinhardt, 1963). Saturation occurred at shallower levels first because vertically channelized flow from the surface produced locally saturated zones. Vertically channelized saturated flow was discussed by deVries and Chow (1973, 1978).

The lower B horizon piezometers rose and fell quickly as rainfall varied, showing a diurnal variation of ± 2 to 5 cm. This variation is only apparent on the piezometer hydrographs during the first half of the experiment when water levels were recorded more frequently. Piezometers 1, 6, 7, 13, and 14 indicated discontinuous saturation. Because these piezometers were shallow, they probably were in the saturated zone only during the higher phase of the diurnal cycle.

The piezometers in the lower B horizon also responded to the decrease in rainfall on the third and fourth day of irrigation (August 10th and 11th) caused by of the partial clogging of the input screen on the pump. This response was shown by falling heads starting late on the 10th and continuing through the 11th. A subsequent rise of up to 0.20 m occurred on the morning of the 12th when irrigation returned to normal.

Five piezometer tips were located at depths greater than 0.30 m into the till. Four of these (A, F, K, and N) demonstrated delayed and damped response. Damping was shown by reduced diurnal variation and smaller response to the lesser rainfall of the 10th and 11th. (The large head drop and recovery in piezometer N on the early 11th remains a mystery. The rise occurred too soon to be indicative of the return to higher rainfall, Figure A5-9). The delayed response in the till is shown by: later initial rise, greater time required to reach steady-state, and longer lag between the temination of rainfall and the initiation of falling head.

Piezometers below the irrigated area responded to rainfall applied to the plot. Piezometer L16 (directly below the outflow trough) initially rose 0.30 m and then indicated a head drop caused by the decrease in rainfall on the 10th and 11th. Piezometer L15 was not read frequently enough to note this reduction. Downslope response was also seen in the piezometers farther below the plot (L6 through L13). The rise in these piezometers was approximately 0.10 m. However, because the rise occurred gradually, time lags could only be resolved to between 70 and 130 hours. Piezometer L8 indicated flowing artesian conditions and therefore a natural discharge area. As an extension to contain the water at its equilibrium level was not properly sealed until the 16th, no head rises were observed in this piezometer.

5.3 Outflow

Outflow from the north, south and combined, tipping buckets is shown in Figure 5-4. The flood wave began 19 1/2 hours after rainfall began with the first peak occurring after 40 1/2 hours. Outflow rose and fell cyclically until steady state was reached after 6 days. This cyclic variation was diurnal and amounted to ± 25% of the mean daily outflow. Based on the steady state period August 13th through 20th, the mean daily maximum occurred at 9:15 hrs (± 15 minutes) with the mean minimum at 19:50 hrs (±

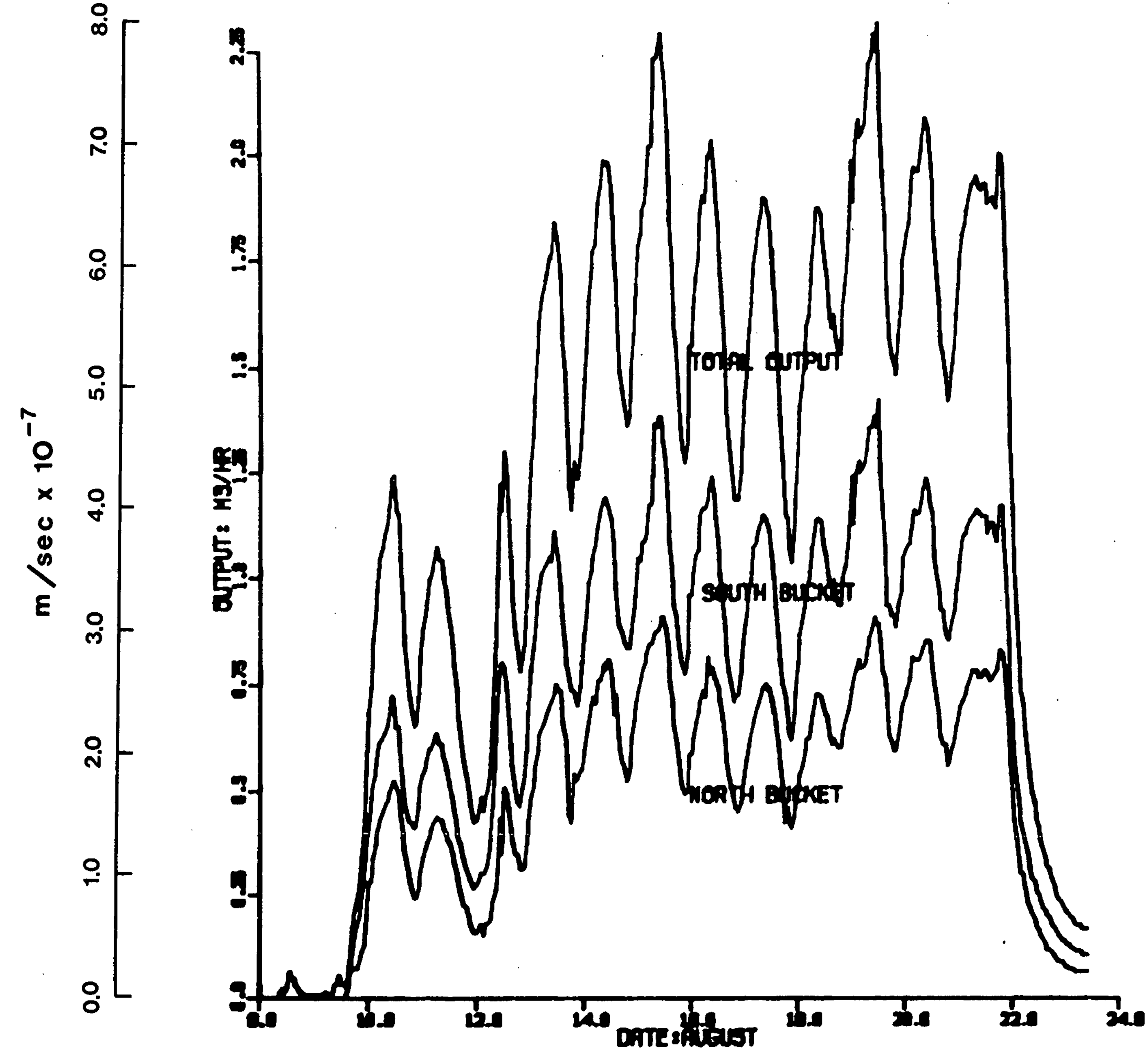

FIGURE 5-4 Outflow measured by the tipping buckets

15 minutes). After precipitation ceased, outflow began to dercease within 1 1/2 to 2 hours. Forty hours later, outflow was 1/12 the rate prevailing before irrigation was shut off.

5.4 Chemistry

The results of the chemical analyses were surprising. It had been hypothesized that high inflow concentrations of PO_4^{3-} and NH_4^+ (10 mg/l and 17 mg/l respectively) and short contact time with the mineral soil would produce high outflow concentrations of these species. Instead, analyses showed no detectable amounts at any sample location (except input). These results include samples taken at in-plot piezometers B and C, which were less than 1.4 m below the surface. It appears that exchange capacities, sesquioxide-phosphate reactions and nitrogen fixing reactions within the forest floor and/or B horizon were more significant than previously expected. Although these species were applied at rates considerably higher than those of Bryck (1977), these findings are still similar to his; all PO_4^{3-} and NH_4^+ were tied up. Because the soil was so adsorptive, this concentration data could not be used to elucidate flow paths.

Chloride results were also surprising. The chloride data for input, in-plot (piezometers B and C), outflow, and below plot (piezometers L15 and L9) are graphed in Figure 5-5. Concentration at the initiation of outflow was negligible. Rising steadily, concentration peaked at approximately 1/2 the input level, more than 2 days after switching to clean water. Analyses of NO_3^- and K^+ revealed similar timings, but with input-outflow concentration ratios that were lower. The two-day-

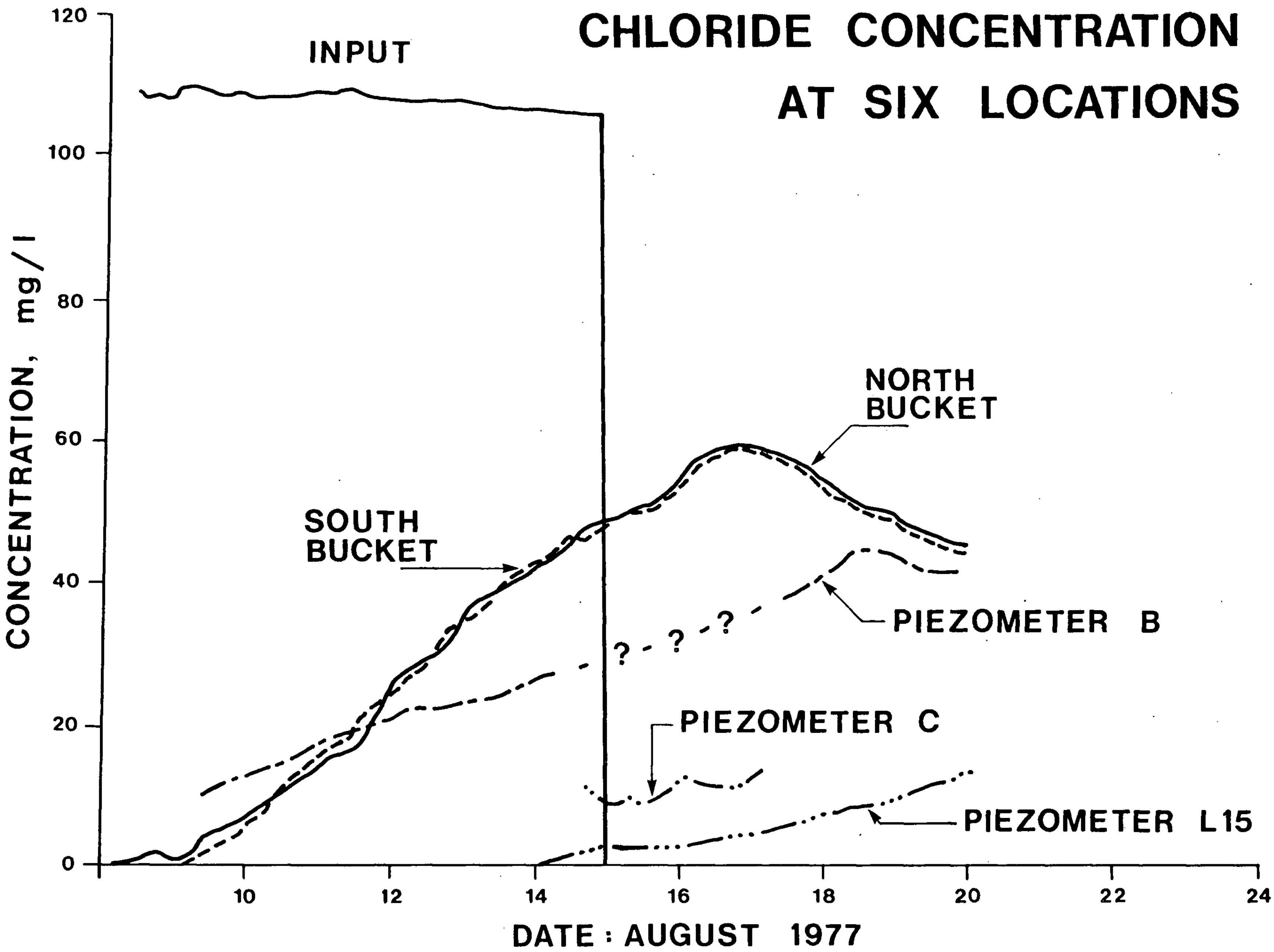

FIGURE 5-5 Chloride Concentration vs. Time at Four Sample Locations: Input, In-plot (Piezometers B and C), Outflow and Below-plot

plus concentration-peak lag and input-outflow concentration ratios of less than one, indicate one or more of the following: the occurrence of flow path reactions transforming input species into unanalysed ones, cation and anion exchange with the soil, dispersion, and / or the mean flow path travel time (average linear velocity). It had been hoped that chloride would act as a non-reacting base level tracer with which to compare phosphate and ammonium. This could not be done and therefore little flow path information was gained.

5.5 Slug Tests

Results of the slug tests are listed in Table 5-3. Hydraulic conductivities ranged from 10^{-7} to 10^{-6} m/s with a mean value around 5×10^{-6} m/s. The lowest conductivity of 10^{-7} m/s was measured in piezometer F located in an area of well defined, more compacted till with a sharp B horizon contact. Upper till conductivities (piezometers B, D, L16, and L15) as well as lower B horizon conductivity (piezometer G) were all approximately 8×10^{-7} m/s. These areas reflect a B horizon to till transition zone instead of a well defined contact. Thus, in transition areas, the hydraulic conductivities in the upper till and the lower B horizon are probably similar but with a slight increase toward the surface.

Piezometer A indicated a conductivity of 9×10^{-6} m/s. This was not accurate as the water level from the injected slug fell at two distinct rates, confirming a suspicion that this piezometer was not sealed just above the intake screen. The calculated value was too high as improper sealing produced an

Table 5-3. Hydraulic conductivities calculated from Hvorslev (1951).

(Where $K = \frac{R^2 \ln(\frac{L}{r})}{2LT_o}$ R = 2.5 cm, L = 15 cm, r = 1.75 cm).

	Piezometer	T_o(min)	K(m/s)	Unit
Slug Test	A	13	9×10^{-6}*	Till & B
	B	330	3×10^{-7}	Till
	D	180	6×10^{-7}	Till
	F	840	1×10^{-7}	Till
	G	140	9×10^{-7}	B
	L16	72	1×10^{-6}	Till
Bail Test	L15	< 120	$> 9 \times 10^{-7}$	Till

* See text.

inflow area larger than that used in the Hvorslev (1951) equation. Therefore, this conductivity is not indicative of the till or the B horizon.

6.0 ANALYSES, INTERPRETATION, AND DISCUSSION

In this chapter, the results from the previous chapter are interpreted and discussed. To begin with, the equation of continuity is used to calculate the non-stormflow ground water input and storage volume of the plot. The water table is then discussed: its nature, position, and role in determining flow paths and gradients. Following this, the bulk hydraulic conductivity of the hillslope plot is calculated and compared with measured values. From all of these results and analyses the mechanism of stormflow generation is then summarized. Finally, two possible objections to generalizing this mechanism to other watersheds are discussed.

6.1 Non-Stormflow Groundwater Input

In previous work by Nagpal and deVries (1976) the term leakage was used to describe the water lost from the soil in the form of flow into the till. Because the approach is being taken that the saturated and unsaturated zones make up one system, the term non-stormflow ground water input will be used instead of leakage. Non-stormflow ground water input is the component of saturated flow which flows into the till and does not exit shortly thereafter.

To calculate the non-stormflow ground water input the equation of continuity can be used:

$$I - (O_s + Q_f) = \Delta S$$

Where: I = Rainfall input

O_s = Outflow at the collection trough

O_t = Outflow into the till

ΔS = The change in storage

At steady state, $\Delta S = 0$ and the flow into the till is equal to the difference between rainfall input and outflow at the collection trough. Steady-state input during the latter part of the rainfall-runoff experiment averaged 2.11 m^3/hr or 7.97×10^{-7} m/s integrated over the total irrigated area of 735 m^2. Outflow during steady state had a mean rate of 1.70 m^3/hr or 6.43×10^{-7} m/s. The difference between these values indicates a loss of 1.54×10^{-7} m/s or less than 20% of input. (This difference can be seen as the area between input and outflow in Figure 6-1). The actual loss was undoubtably less than this figure as piezometers A, F, and N were slowly rising during this period, indicating that some water was still going into storage. Non-storm ground water flow out of the plot was confirmed by piezometric response below the irrigated area.

There are two reasons why this calculated 20% loss is significantly less than the 75% reported by Nagpal and deVries . The first is that rainfall rates were calculated differently. Nagpal and deVries, calculated rainfall at 0.6 cm/hr as opposed to the mean value of 0.287 cm/hr for the latest experiment. Their figure is similar to the highest rates measured in exposed (i.e. no vegetation cover) rain gauges in this latest experiment where rainfall rates varied from 0.07 cm/hr to 0.72 cm/hr. Thus, the use of Theissen weighted polygons and representative placement of rain gauges gave a rainfall rate that was less than half of that estimated previously. This helped produce a non-stormflow ground water component that was less than five times

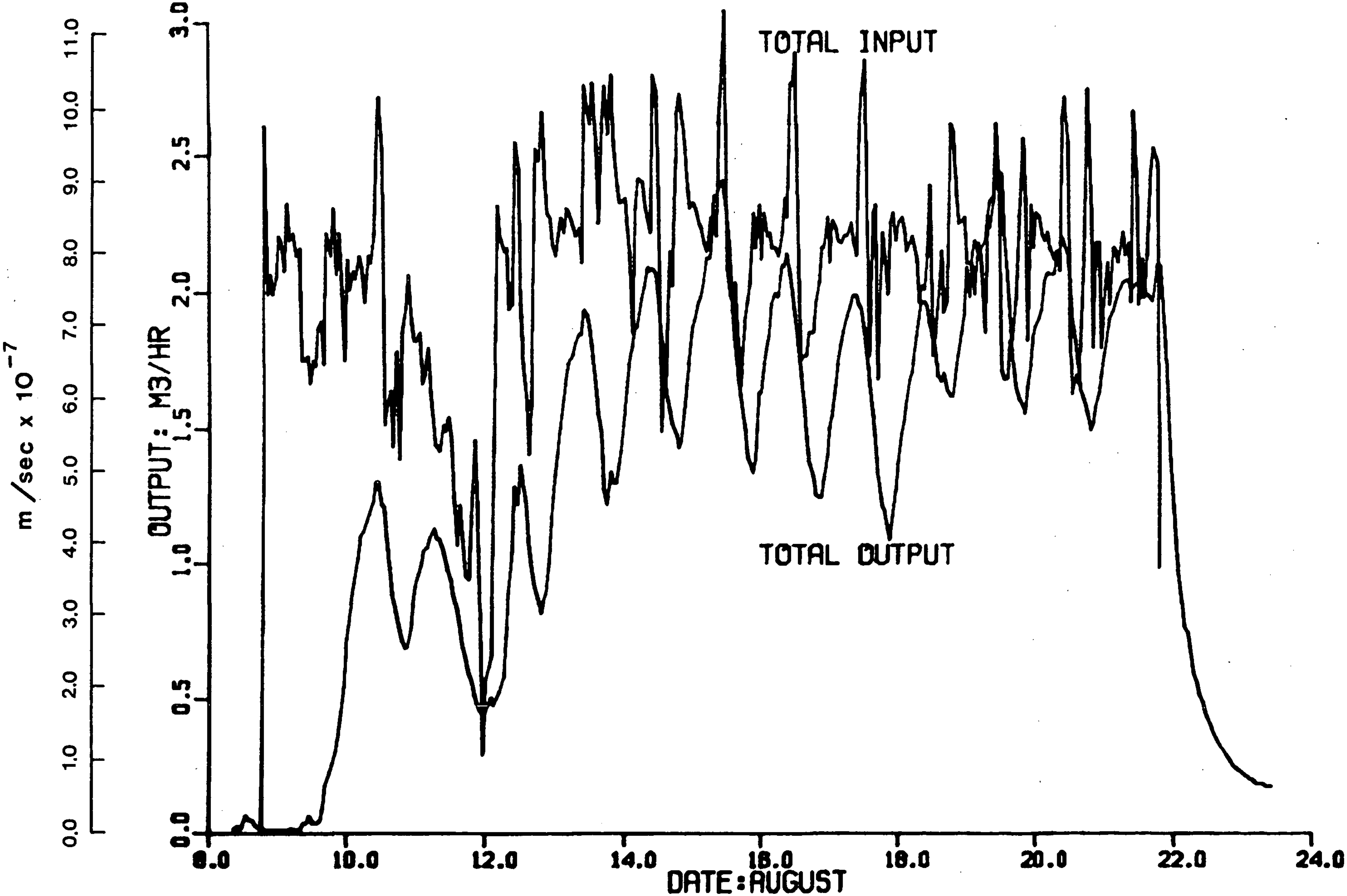

FIGURE 6-1 Total input and output of the experimental plot

smaller than that calculated by Nagpal and deVries.

The lower rainfall rate was supported by a comparison of rainfall and pumping rates. The rainfall rate of 0.287 cm/hr is equivalent to a no-loss pumping rate of 2.11 m^3/hr. The actual pumping rate averaged 3.11 m^3/hr. The 32% difference was supported by observations of sprinkler overspray and clouds of mist advected out of the plot during irrigation. The higher rainfall rate of 0.60 cm/hr would have required a pumping rate of 4.41 m^3/hr without taking into account advective and overspray losses. Thus, if the higher rainfall rate was accurate, the pumping rate would have to have been at least 42% higher than that measured in the latest experiment. Since the same constant-rate irrigation system was used for both experiments it is unlikely that the higher rate is valid.

A second reason for the discrepancy in the ground water input values is that Nagpal and deVries' calculations were not based on true steady-state. It can now be seen that the leveling off of outflow which they attributed to the onset of steady-state conditions was really just an outflow diurnal peak. Had they continued irrigation, outflow would have increased to a higher rate through several more diurnal cycles. Thus, some of the difference between input and outflow which they attributed to leakage from the soil was actually due to water going into storage. For this reason as well as their overestimated rainfall rate, Nagpal and deVries' estimation of the non-stormflow ground water input was probably much too high. The 20% loss rate suggested here is more likely to be correct.

The diurnal variation in both input and outflow coupled with a much lower non-stormflow ground water input rate indicated that the previous interpretation of water breaking through imperfections in the compacted till is not necessary to explain the loss from the system. It is unlikely that a temporally and spatially discontinuous process such as hydrologic breakthrough would have occurred at the same time each day. It is more likely that the variation in outflow was directly caused by the variation in input as both of these volumes and timings were similar (Figure 6-1). Synchonous timings were also seen in hillslope piezometers. Thus, the variation in input caused a variation in piezometric levels which in turn caused a variation in outflow.

6.2 Soil Moisture Storage

The soil moisture storage volume was calculated using the equation of continuity and the assumption that the difference between transient and steady state flow rates into the till was not significant. This assumption seems reasonable because the increased flow into the low conductivity till (induced by transient gradients) should be proportionally much smaller than either outflow or rainfall input. Thus, for example, if the estimation of flow into the till were off by 50%, the error in storage would be less than 10%.

The total rainfall during the 6 day transient period was 302.1 m^3 while outflow measured 145.6 m^3. Flow into the till, at 20% of rainfall input, equals 60.4m^3.

Then:

$I - (O_s + O_t) = \Delta S$ Symbols defined above

$\Delta S = 91.2$ m^3 or

$= 0.131$ m water depth equivalent

Based on an estimated porosity of 0.4 for the lower B horizon and the assumption that the coarse texture of the B horizon allowed for little unsaturated storage, This S value produced a water table rise of 0.38 m. This rise was consistant with that indicated by the in-plot piezometers.

6.3 The Water Table

6.3.1 Configuration of the Water Table

Water going into storage caused either a continuous rise in the saturated zone or the formation of a perched water table. The piezometric data did not always indicate where each of these situations predominated. In places that were saturated both above and below the B horizon - till contact before irrigation began (such as near piezometers A, B, C, D, and N), the saturated zone remained vertically continuous throughout the experiment. Initial piezometer response at these locations was generally faster than in places where piezometers were initially dry. Such behavior is consistent with classical infiltration theory (Rubin and Steinhardt, 1963) as the water table rose, not because the infiltration rate was higher than the saturated conductivity but because storage was being filled.

In areas where piezometers were initially dry, the water table configuration was not as clear. Either the filling of storage caused a vertically continuous saturated zone to rise or the infiltration rate of 7 x 10^{-7} m/s led to the formation of a perched water table above the surface of the till ($K = 10^{-7}$ m/s). The initial piezometric response time-lag could not be used to distinguish between a perched or continuous water table. It was not possible to tell whether time-lags were a function of: flow path-length, a function of the filling of storage above the till to form a perched saturated zone, or a function of the depth to the saturated zone. Where the B horizon - till contact was well defined, it is most likely that a perched water table existed (as seen at other locations, see 3.3.2). A gradational contact probably hosted a vertically continuous saturated zone because the lower conductivity of this zone would produce slow drainage. Thus, water would be in storage long enough for a saturated front to move downward, eliminating any underlying unsaturated region. Since both well-defined and gradational contacts existed within the plot, it is likely that both types of water table configurations occurred.

Fortunately, the nature of the saturated zone did not control where subsurface stormflow occurred. In both perched and continuous water table situations, most stormflow traveled through the high conductivity B horizon because the conductivity contrast with the till and B horizon - till transition zone was 2 to 3 orders of magnitude. The low conductivity of the till and transition zones kept stormflow contributions from this region to a minimum. Thus, it was not important to the overall volume of stormflow whether the underlying low conductivity zones were saturated or not as contributions would have been small in either situation.

6.3.2 Role of the Water Table

The role of the water table was revealed by the piezometric and outflow data. The piezometers indicated that the overall hydraulic gradient remained approximately constant while water levels and outflow varied diurnally. Four piezometers representative of the lower, middle, and upper slope demonstrated that within the resolution of four readings per day, diurnal timing was independent of hillslope position (Table 6-1). Piezometer variation at each of these locations was also approximately the same (± 5 to 10 cm). Thus, the water table rose and fell parallel to the overall hillslope. The parallel response of the water table and the relative thinness of the zone through which most saturated stormflow occurred produced flow paths that were approximately parallel to the hillslope and gradients that were similar to that of the overall hillslope (Figure 6-2). These gradients remained approximately constant (equal to the tangent of the hillslope) because local gradient differences from variations in water table position were unimportant relative to the overall gradient of the hillslope. Thus, the main component of the hydraulic gradient was gravitational.

Table 6-1. Maximums occurred in late morning (8:00 to 10:00) with minumum at early evening (18:00). Values on the 12th are not representative because of rainfall reduction.

Piezometer	2		8		11		L	
Position	Lower		Lower		Middle		Upper	
August	Min	Max	Min	Max	Min	Max	Min	Max
9	Rising		12:00	8:00	17:00	7:00	Rising	
10	17:00	11:00	18:00	8:00	17:00	8:00	17:00	3:00
11	1:00	6:00	2:00	7:00	19:00	7:00	24:00	7:00
12	19:00	12:00	20:00	13:00	20:00	8:00	19:00	13:00
13	18:00	6:00	Lost		19:00	7:00	18:00	7:00
14	18:00	10:00	18:00	11:00	19:00	11:00	18:00	11:00

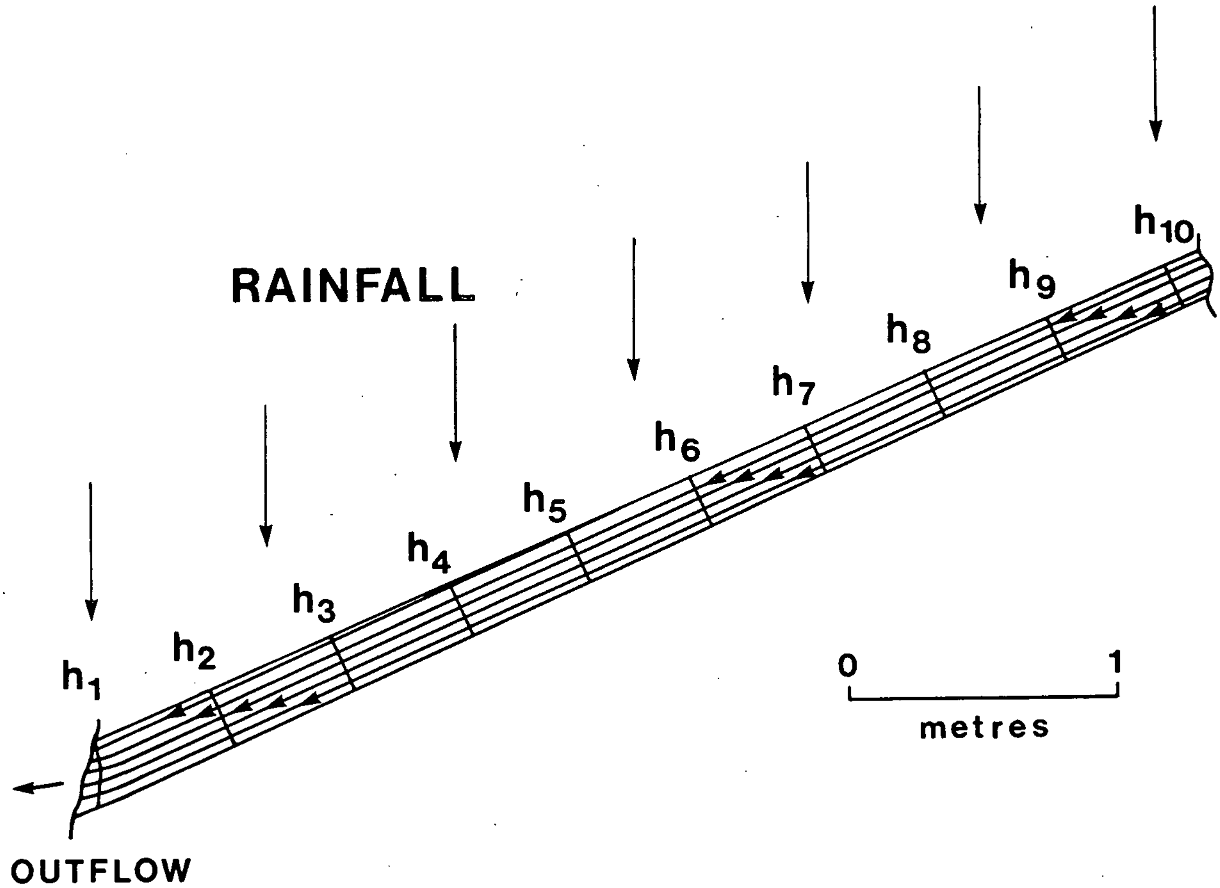

FIGURE 6-2 Schematic Flownet of part of the Hillslope
Flow is parallel to the hillslope because the saturated stormflow zone is relatively thin compared to the total hillslope length. Gradients are primarily gravitational.

Classical hillslope behavior, where the magnitude of water table fluctuations decrease with distance from the stream, was not observed. There are several possible reasons for this. First, the high-conductivity B horizon was probably able to accommodate the greater flow nearer the stream bank with only a slightly greater water table rise than in upslope regions. Second, this rise was not seen because the large heterogeneity of the B horizon produced water table rises that were extremely variable for a given distance from the stream bank. Thus, the water level indicated by each piezometer may not have been indicative of the mean water table height for its distance from the stream. A slightly larger diurnal variation near the stream bank would probably have been seen with a larger piezometer array.

Gradients and saturated conductivities remained approximately constant with time, indicating that variations in outflow were caused by variations in the saturated cross-sectional area available for stormflow . Within this area, delineated by the water table on top and by the low conductivity till and B horizon -till transition zones on the bottom, a stormflow area variation of ± 10 to ± 20% caused an increase and decrease in outflow of ± 25%. The smaller variation in saturated flow area causing a large variation in outflow indicated that a greater proportion of the stormwater was flowing in the upper portion of the saturated zone. This conclusion was supported by low conductivity measurements in the till and lower B horizon.

Parallel water table response and proportional outflow were also observed during the terminal phase of the experiment. After irrigation was shut off most piezometric levels began to fall in less than two hours as did the outflow rate. For the following few hours water levels fell, maintaining an approximately constant gradient, but with decreasing cross-sectional flow area and a corresponding decrease in outflow rate. After this time, runoff decreased while water levels fell only slightly supporting the conclusion that higher conductivity and a greater proportion of flow occurred in the upper part of the saturated zone.

The importance of the position of the water table in generating outflow was clearly demonstrated in its behavior just before outflow began. Piezometer L16, just below the collection trough, began to rise 9 to 10 hours before any runoff was produced. Outflow commenced only after the water table rose and intersected the surface of the stream bank as evidenced by water levels in piezometers L16 and A (Figure 6-3). Before outflow began, water was going into storage causing either a water table rise in the already saturated B horizon -till transition zones or the formation of a perched water table where the B horizon - till transition was initially unsaturated. After this time the rising water table produced an increasing amount of stormflow as the zone of saturation rose into the organic rich zones of the lower B horizon. Outflow was at a maximum when the cross-sectional area through which it flowed was also at a maximum. Thus, the position of the water table directly controlled the rate of outflow.

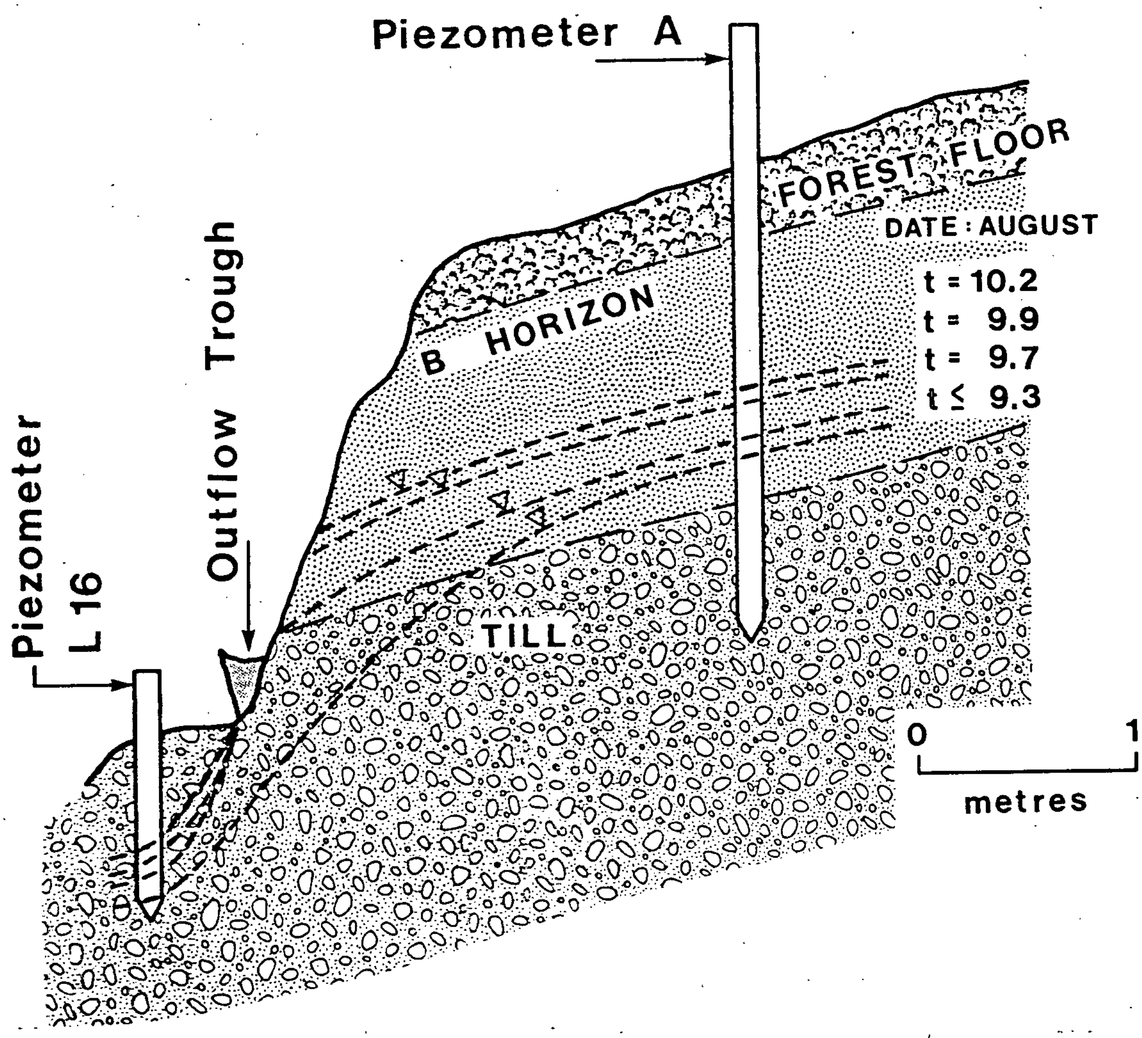

FIGURE 6-3 Water Table Position Around the Initation of Outflow

6.4 Hydraulic Conductivities

6.4.1 Calculated Conductivities

After it was established that stormflow was parallel to the hillslope, the bulk conductivity of the plot was calculated using Darcy's law. This calculation was made for a zone a short distance in from the seepage face because, at the face, exit effects and the effects of the steeper seepage face angle produced gradients and flow paths that were hard to estimate.

Darcy's law can be written:

$$Q/A = K \text{ grad } h$$

Where: Q = outflow (L^3/T)

A = cross-sectional flow area (L^2)

K = saturated hydraulic conductivity (L/T)

grad h = hydraulic gradient (L/L)

The mean steady-state outflow rate of 2.11 m^3/hr was calculated for Q. A cross-sectional flow area of 15 m^2 was estimated using the plot width of 30 m and a saturated depth to the low conductivity layer of 1/2 m. This depth of the stormflow area was based on piezometer rises recorded near the stream bank and on visual observations at the seepage face during steady state. The hillslope gradient of 0.4 was used for the hydraulic gradient, a value that is realistic for all points throughout the hillslope except for those right at the seepage face.

With these values, the bulk hydraulic conductivity of the research plot was calculated at 8×10^{-5} m/s. This conductivity is probably a minimum as the saturated flow depth of 1/2 m was the upper limit of values indicated by the in-plot piezometers. The actual flow depth was probably less because most stormflow occurred in the upper part of the saturated zone. Thus, part of this 1/2 m included area through which only a small proportion of stormflow traveled. Therefore, the bulk conductivity of the stormflow transmission zone was probably somewhat greater than the calculated value of 8×10^{-5} m/s.

6.4.2 Comparison of Conductivities

The calculated bulk hydraulic conductivity indicated that most storm runoff flowed through a unit with conductivities three orders of magnitude higher than those measured for the till and lower B horizon. Yet, examination of the seepage face and piezometer rises revealed that most flow did occur in the lower part of the B horizon. This paradox is most likely explained by the presence of organic zones, mostly live and decayed root material which produced a much higher overall conductivity. Water flowing through the B horizon passed preferentially through these zones, which, having a very high conductivity, gave a bulk conductivity 2 to 3 orders of magnitude larger than the mineral matrix of the lower B horizon alone.

Most runoff exited from the stream bank via these root channels. The small proportion of outflow that came directly from the soil matrix did not indicate that the matrix was uninvolved in stormflow generation nor did it indicate that the organic pathways were interconnected channels. A schematic flow net (Figure 6-4) demonstrates that the high conductivity contrast between the root channels and the soil matrix is sufficient to cause a major portion of flow to exit via the organic channels. Up the hillslope, water traveled between adjacent root channels via the saturated matrix of the lower B horizon. Thus, the bulk conductivity of the soil was determined mostly by the concentration of high conductivity organic channels.

6.5 Mechanism of Stormflow Generation

On the basis of this study, and the work of deVries and Chow (1973, 1978) and Nagpal and deVries (1976), the following mechanism of stormflow generation is envisioned:

Rain hits the ground and flows along leaves, branches, logs, rocks, etc. into and through the forest floor. This flow is not unsaturated as the porosity and conductivity of the forest floor might suggest. Rather, flow is locally saturated in vertical channels because of the extremely open nature of the forest floor and the concentrating effects of logs, branches, etc. These concentrating elements allow for the formation of localized free water which can then enter the large open pores in the soil. The open nature and high conductivity of the forest floor also preclude ponding and Hortonian overland flow except

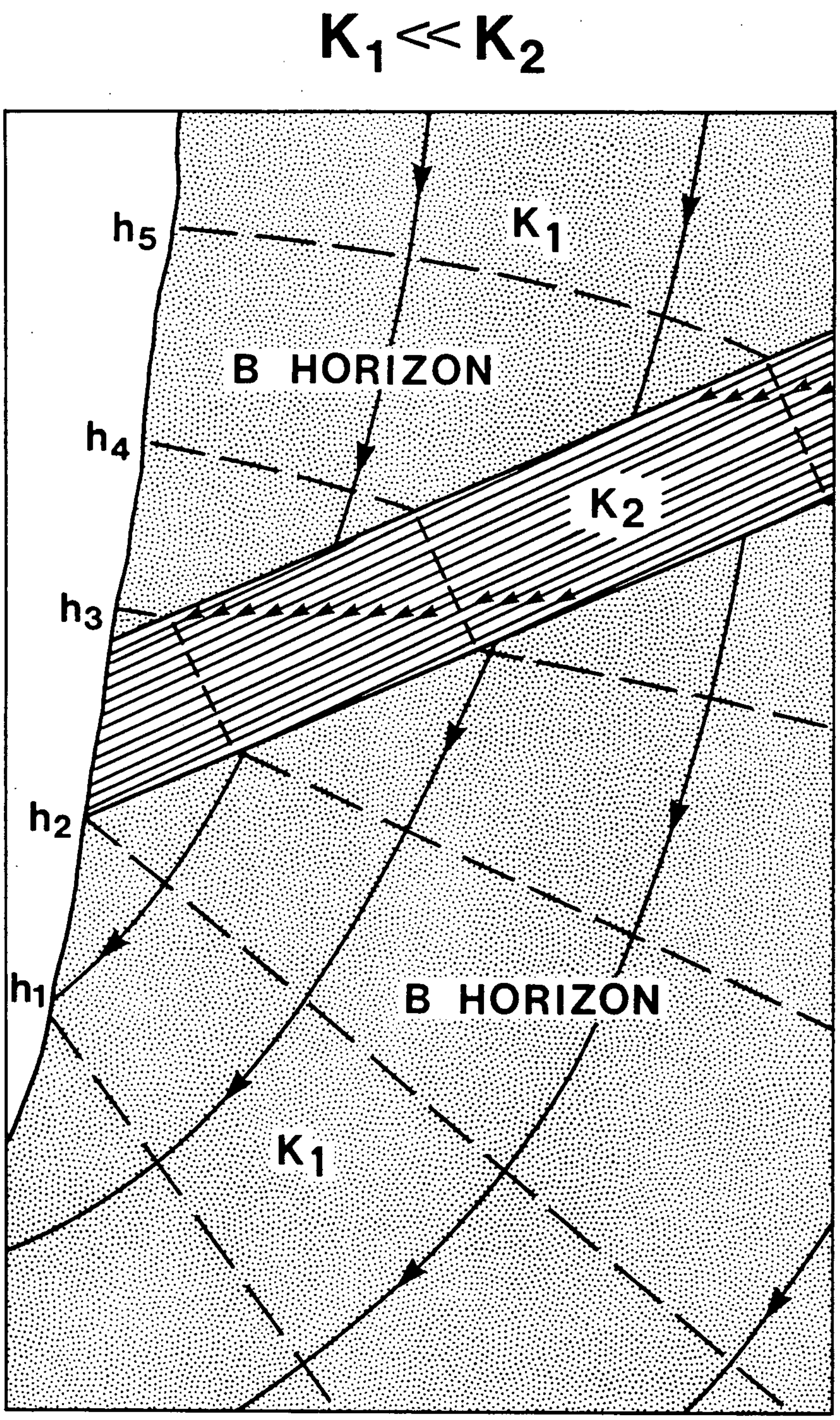

FIGURE 6-4 Schematic Flownet of the Stream Bank showing prefered Path Through High-conductivity Root Zone

in locally disturbed areas.

Water flows downward through the forest floor until it encounters the B horizon. It then enters and continues to flow downward through locally saturated pathways, primarily through high conductivity organic root zones. Vertically channelized, saturated flow predominates because the lower saturated conductivity of the soil matrix (relative to the organic channels) allows only a small amount of outward flow into the matrix.

Because vertical flow through the B horizon is concentrated into high-conductivity, saturated regions, flow downward to the water table or, if is encountered first, to the low conductivity till, is rapid. In the lower part of the B horizon, this downward moving water fills storage and causes either the water table to rise or the formation of a perched water table where the lower conductivity of the till impedes downward flow.

As water goes into storage, the water table rises. During this period only a small amount of water flows out at the artificial stream bank because the water table is below the surface of the collection trough. Any outflow which does occur during this period is probably due to a locally perched water table near the stream bank.

Major outflow begins when the water table intersects the surface of the stream bank. At this point, most subsurface storm water is flowing through the high conductivity B horizon. The high conductivity is due to the concentration of organic zones, primarily live and decayed roots. At the stream bank most outflow occurs through these root channels because the contrast in conductivity between the channels and the soil matrix is several orders of magnitude. Only a small proportion of runoff exits through the saturated matrix.

Variation in outflow is a direct function of change in the cross-sectional area through which storm water flows and not of changes in hydraulic gradient. The water table rises and falls parallel to the overall hillslope. Thus, the hydraulic gradient remains approximately constant. Only the vertical cross-sectional saturated flow area changes.

Once outflow has begun, short lag-times between input and outflow variations are possible because short flow paths from the surface to the saturated zone produce fast water table response. A variation in the water table position causes an almost instantaneous change in outflow. During this period the hillslope-stream system is very sensitive to rainfall variations.

During outflow, less than 20% of the infiltrated rainfall enters a deeper, non-stormflow, ground water system. This water moves through the till and possibly the fractured bedrock. Flowing downward and out of the research plot, this water undoubtedly contributes to the base flow of the natural stream below the plot.

With the termination of rainfall, outflow falls off rapidly as the water table drops. This drop does not change the overall hydraulic gradient appreciably as the water table continues to remain approximately parallel to the hillslope. Only the vertical cross-sectional flow area of the saturated high conductivity B horizon varies. Thus, as the water table falls, the decrease in flow area causes a decrease in the outflow rate.

After a few hours, the rate of outflow continues to fall, but with only a small accompanying drop in water table position. During this period, the water table falls into the part of the lower B horizon where root channel concentration is lower.

After the water table drops below the surface of the seepage face, outflow is produced only where the water table is locally perched near the stream bank. Baseflow is not sustained by saturated flow through the till because the water table is below the collection trough. Low outflow continues (from the decreasingly smaller perched water table regions near the bank) until the next rainfall event occurs.

The cycle then repeats.

6.6 Generalization of the Results

There are two objections that could be raised towards generalizing this mechanism of stormflow generation from the research plot to surrounding watersheds. The first of these pertains to the representativeness of the man-made experimental system in comparison to natural systems. The second concerns the observed lag-times. I will try to show that neither of these objections is serious.

It was demonstrated in a previous part of this study that the hydrogeologic units of the research plot were similar in type and thickness to surrounding areas. What remains open to question is whether the hillslope-stream system and corresponding stormflow generation mechanism are typical of nearby watersheds.

Comparison of the natural stream below the research plot with the artificial stream bank and collection trough shows several differences. In contrast with the artificial stream, the piezometric data near the natural stream revealed strong discharge gradients as well as a water table near the surface. In such an area, stormflow generation by the Dunne and Black mechanism (1970a, b) could occur. In the research plot however, the greater depth to the saturated zone and the higher bulk conductivity of the B horizon make this an unlikely mechanism. Dunne and Black reported a major proportion of runoff generated near the stream channel with some upslope areas contributing little, if any, to direct runoff. Therefore, the possibility exists that in a complete basin, larger than the experimental plot, runoff generated by the Dunne and Black mechanism might predominate. However, this possibility is small. An examination of nearby hillslopes revealed that the proportion of wet, near channel source areas was much too small to account for rainfall-runoff ratios reported by Cheng (1975) for local basins. Thus, it is probable that the Dunne and Black mechanism is only locally significant in B.C. Coast Mountain basins.

Examination of upland watersheds in the vicinity of the research plot revealed aspects of the experimental plot that were physically similar to the natural systems. Typically, steep slopes fed incised stream channels which were dry during the summer months. The area around these channels was not marshy nor was the water table at or near the surface. The only major difference between the experimental site and surrounding upland basins was that in the experimental site the till surface was exposed at the stream bank while in the natural system this was rarely the case. The exposure of the till and the position of the artificial stream channel change the initial outflow response time lag. This is discussed further, below.

Input-outflow lags constitute a second possible objection to generalizing the experimental results. The initial response lag of 19 hours appears to be too long to be representative of lags in natural systems. For example, Dunne and Black reported stream response within several minutes after the initiation of rainfall. However, these lag-times were for summer storms with stream response due only to direct precipitation into the stream channel. Undoubtedly, had the roof over the stream bank and collection trough been removed, short lag times of a similar nature would also have been observed for the research plot.

Response times reported by Cheng (1975) are more representative of natural B.C. South Coast Mountain basins. In studying 33 natural rainfall events which occurred in the Jamison Creek basin located in the Seymour Watershed , Cheng noted initial reponse lag-times of 5 to 15 hours. This basin had geologic (and probably hydrologic) characteristics similar to the watershed used for this study. Shorter lags for a larger basin seem to indicate that the research plot is not representative of the local situation. For several reasons these shorter lags are not perceived as a problem.

One of these reasons is that in the mechanism operating in the research plot, the initial response lag is due to water going into storage. Therefore, this lag is a function of antecedent moisture and thickness of the unsaturated units. The experiment at the research plot was conducted after two months of no rain. Cheng's data were for summer rainfall events following an unusually wet period (N. Penny, pers. comm. 1978) as well as for events during the wetter months of fall and winter. For a more direct comparison with Cheng, it is possible to obtain qualitative wet-antecedent input-outflow response lags by considering the sensitivity of the experimental hillslope once outflow had begun. Refering to Figure 6-1, over the period August 15-17 (when the effect is most clearly seen) input-outflow maximum lags were less than 1 hour. Input-outflow minimum lags were from 3 to 7 hours. Thus, once moisture storage requirements had been satisfied, response times were much shorter and as a result, consistent with Cheng. As the thicknesses of the geologic units in the Jamison Creek basin were not given, it is also possible that the longer lag observed in the research plot was due to deeper soil profiles.

The difference in size between Cheng's basin and the research plot is not a major factor in the difference in response lags because outflow is generated by a water table rising parallel to the hillslope in a synchronous manner. Lag time is not a function of individual particle flow path travel times but of moisture content, depth to saturation, and flow paths from the surface. These factors vary with, but are not directly controlled by, basin size. Cheng's larger basin does not necessarily require longer lag-times than the smaller experimental watershed used in this study.

A final reason for the longer response lag in the research plot was the artificial nature of the collection trough. The trough sat on the surface of the exposed till with the water table well below it. Because the saturated zone below the runoff trough started to rise 9 to 10 hours before the initiation of outflow, it is possible that if the artificial stream were topographically lower (and more representative of a natural stream with the water table coincident with the stream) outflow would have commenced sooner. Thus, initial response lag-times could have been as short as 13 1/2 hours (initial response time of piezometer L16), a figure in the range found by Cheng.

It is therefore concluded that even though the experimental hillslope-stream system is not natural, the mechanism of stormflow generation described in this report can be generalized (cautiously) to surrounding watersheds.

7.0 SUMMARY AND CONCLUSIONS

The objectives of this study were three-fold. The first was to examine the underlying glacial till and to determine its physical characteristics, spatial distribution and hydrologic behavior. Emphasis was placed on the mechanism of hydrologic breakthrough proposed by Nagpal and deVries (1976) and the role of the till in controlling leakage out of the soil system. The second was to investigate the mechanisms of stormflow generation operating within the experimental plot and to examine the role of organic channels. The third objective was to establish whether this mechanism could be generalized to similar watersheds. All three of these objectives have been met.

It was determined that the underlying Vashon till has an average particle size distribution of 83.9% sand, 8.4% silt, and 7.7% clay. It can either be hard and well compacted or soft and much looser with the difference attributable to the degree of weathering and not to genetic differences. The spatial distribution of the till is variable. Till is present in 60% of the area near the research plot at depths of 0.15 to more than 1.5 m. The hydraulic conductivity of the till is 10^{-6} to 10^{-7} m/s. Leakage rates through the till are less than 20% of input. The proposed mechanism of hydraulic breakthrough is unnecessary to account for the observed outflows. The variation in outflow is attributable to diurnal variation.

Stormflow is generated when infiltrated rainfall causes a water table rise into the high conductivity lower B horizon . The bulk conductivity of the lower B horizon is high due to the presence of numerous high conductivity zones consisting of live and decayed roots. Most storm water presumably travels through these root channels because the conductivity of the lower B horizon matrix as determined by piezometer tests is very low, 10^{-6} to 10^{-7} m/s. A major proportion of the stormflow exits at the stream bank from these root channels because of the conductivity contrast. Only a small amount of water exits from the saturated matrix of the lower B horizon.

Outflow is controlled by the position of the water table. Vertical fluctuations in response to rainfall causes changes in the cross-sectional flow area perpendicular to flow. The area available for saturated flow controls the outflow rate as the hydraulic gradients stay approximately equal to the hillslope gradient. Outflow ceases when the water table drops below the contact between the B horizon and the lower conductivity till.

This mechanism of stormflow generation can be generalized (cautiously) to similar watersheds. A geologic study demonstrated that the hydrogeologic units of the research plot are representative of the surrounding area. The outflow trough, however, is different from a natural stream. This difference probably causes a delay in inital outflow response which helps to explain why initial outflow lag-times at the experimental plot are longer than those of a nearby watershed. It is concluded that these longer lags do not limit generalization because most of this lag time is a function of soil moisture storage requirements and position of the outflow trough. Once moisture requirements have been met and outflow has commenced, the hillslope plot is just as sensitive to input variations as are natural hillslopes. It is concluded that the experimental plot in this wetter state yields lag-times that are consistent with local catchments.

This report has attempted to explain stormflow generation in a southwest Coast Mountain environment. Many questions have arisen during the course of this research to join those questions that have yet to be asked. I hope that this report will be used as a stepping stone toward answering these questions and that a more complete understanding of hillslope hydrologic processes will result.

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APPENDEX: PIEZOMETER HYDROGRAPHS

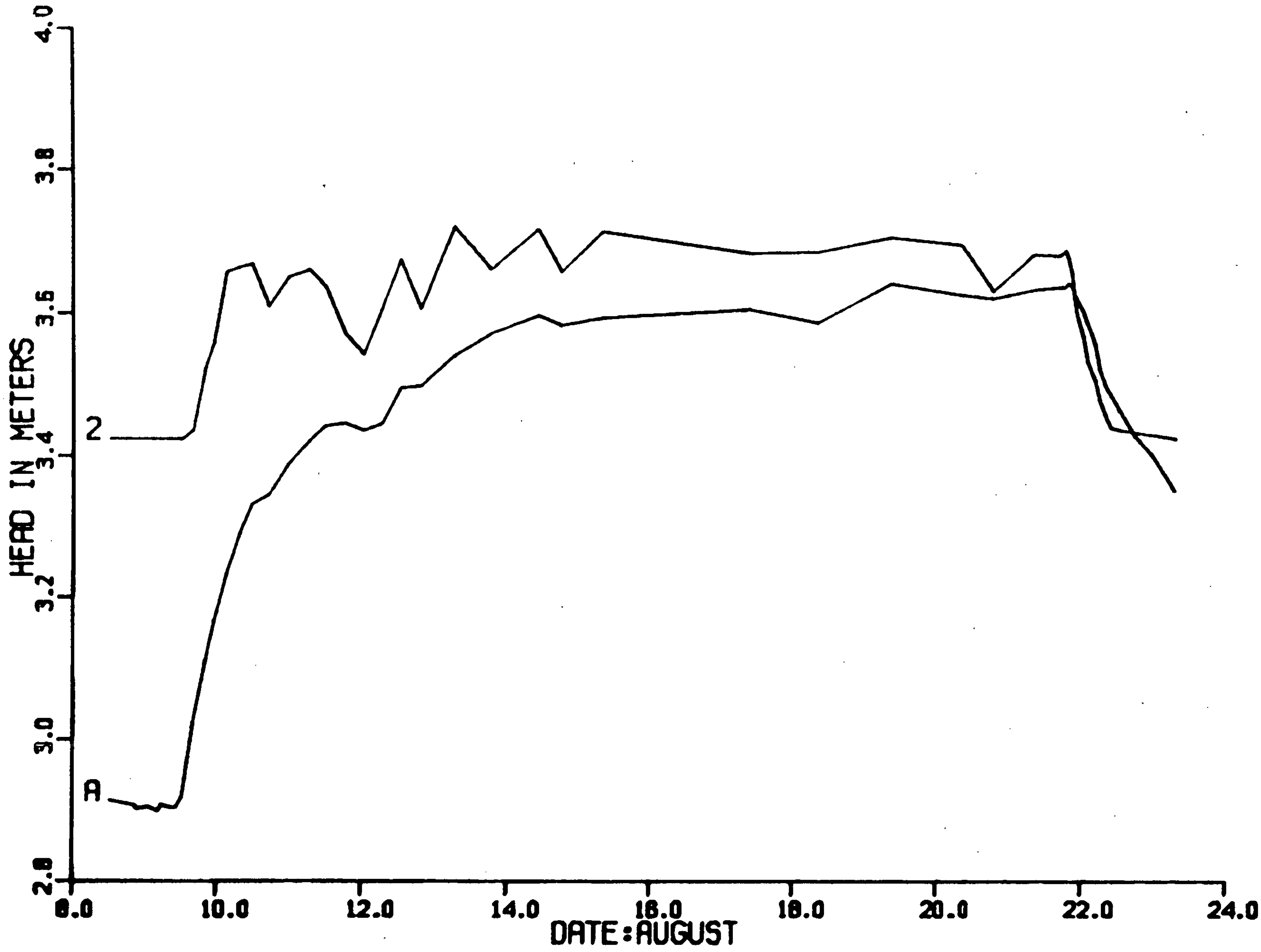

FIGURE A5-1 Hydrograph for piezometers 2 and A

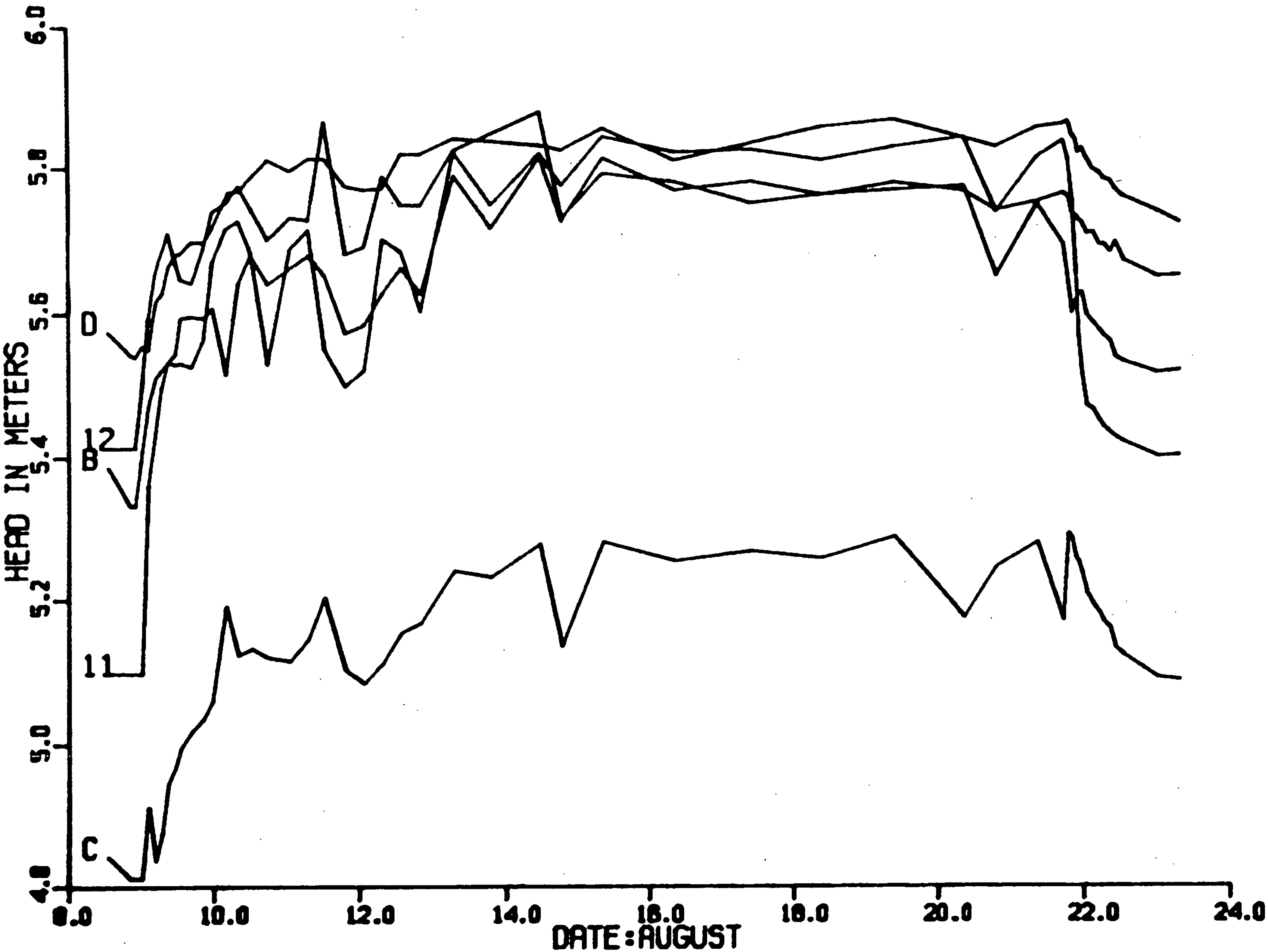

FIGURE A5-2 Hydrograph for piezometers 11, 12, B, C, and D

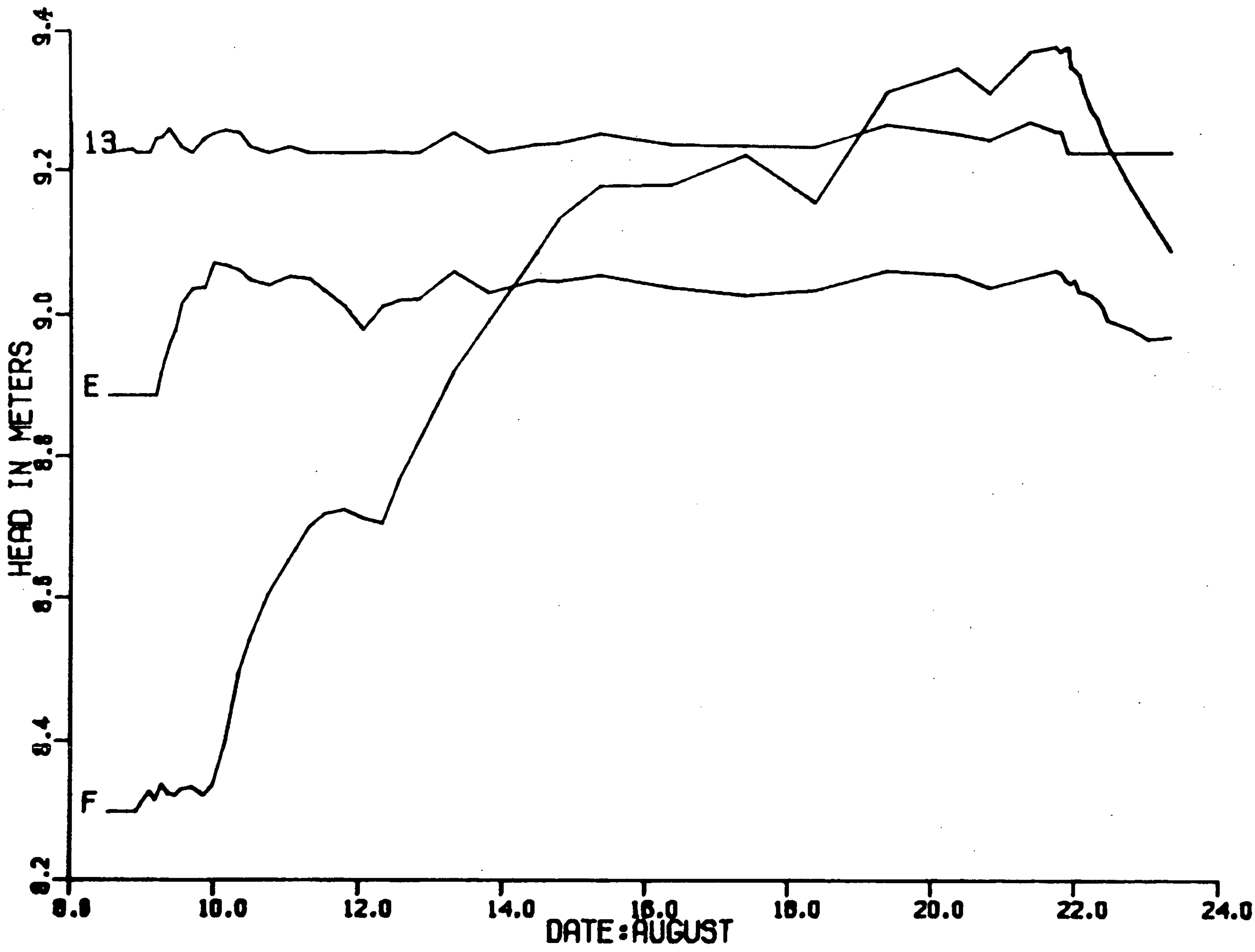

FIGURE A5-3 Hydrograph for piezometers 13, E and F

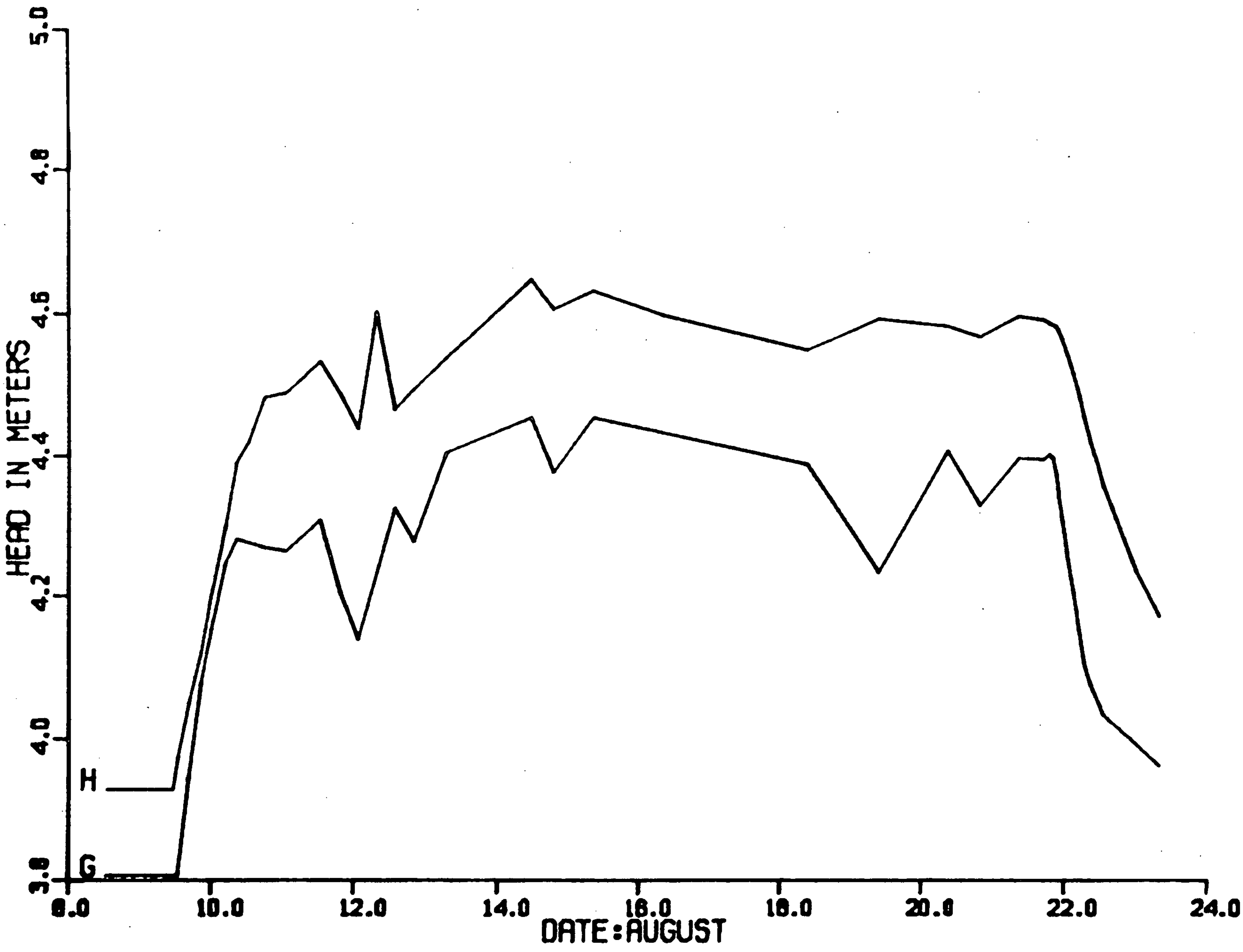

FIGURE A5-4 Hydrograph for piezometers G and H

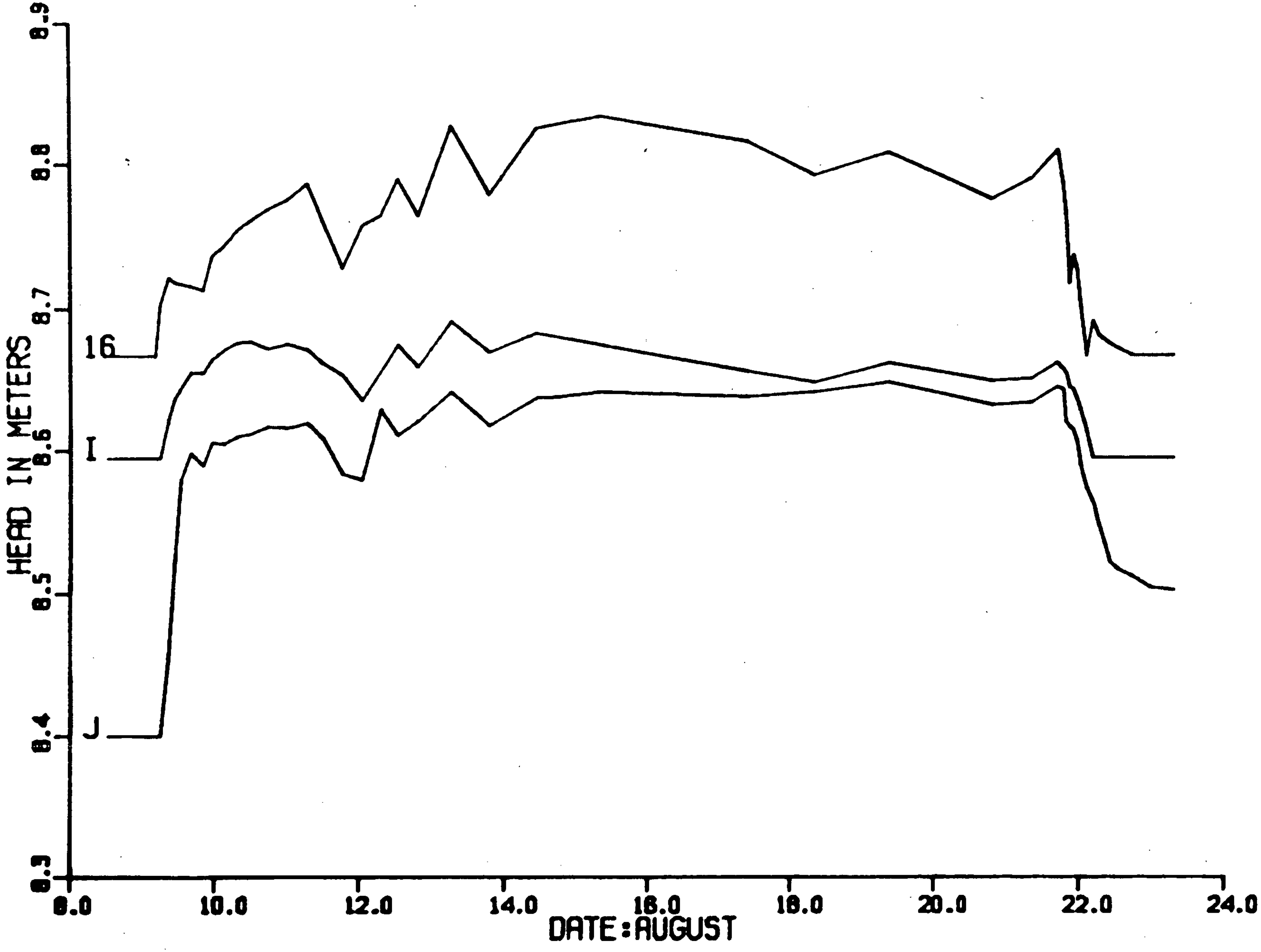

FIGURE A5-5 Hydrograph for piezometers 16, I and J

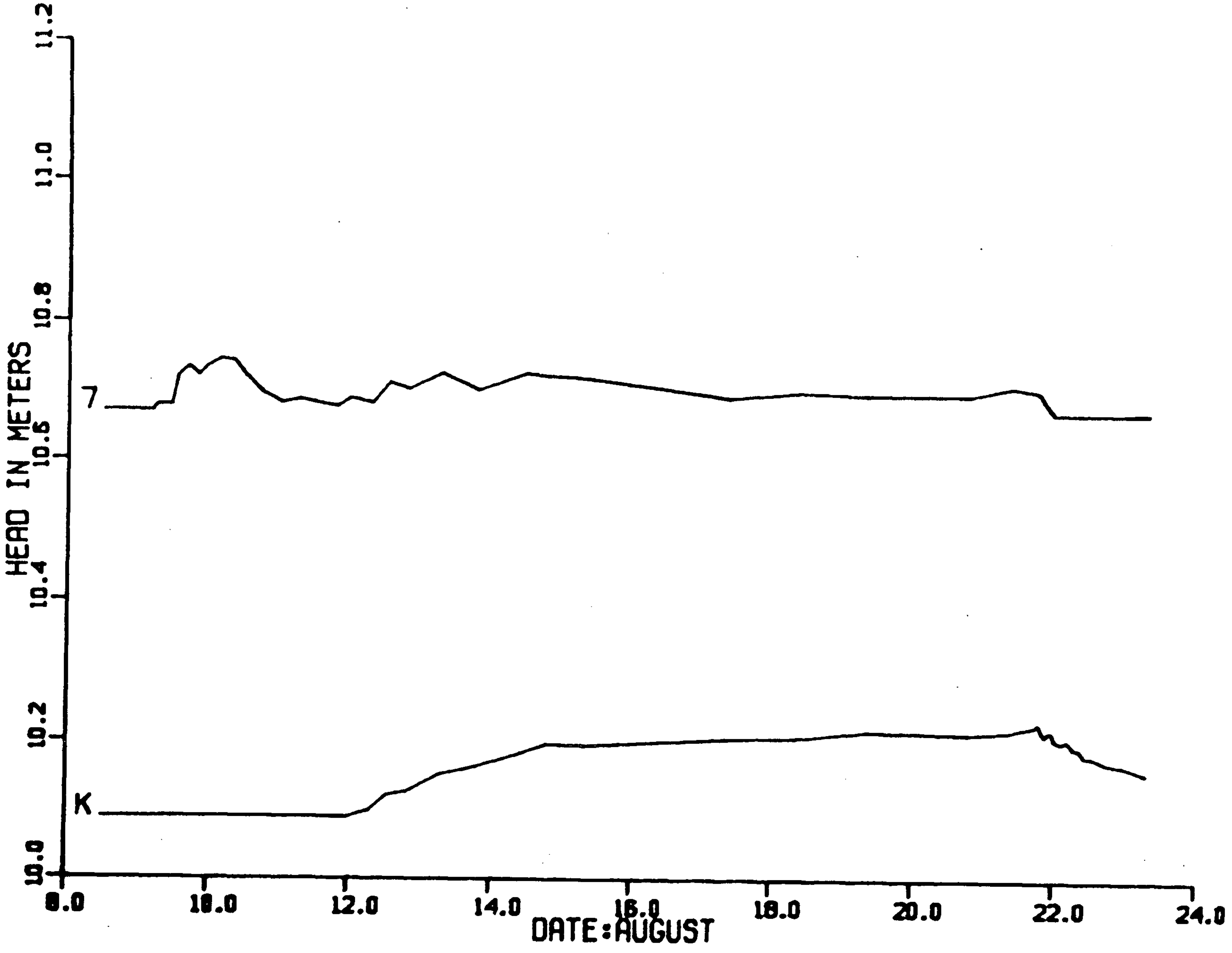

FIGURE A5-6 Hydrograph for piezometers 7 and K

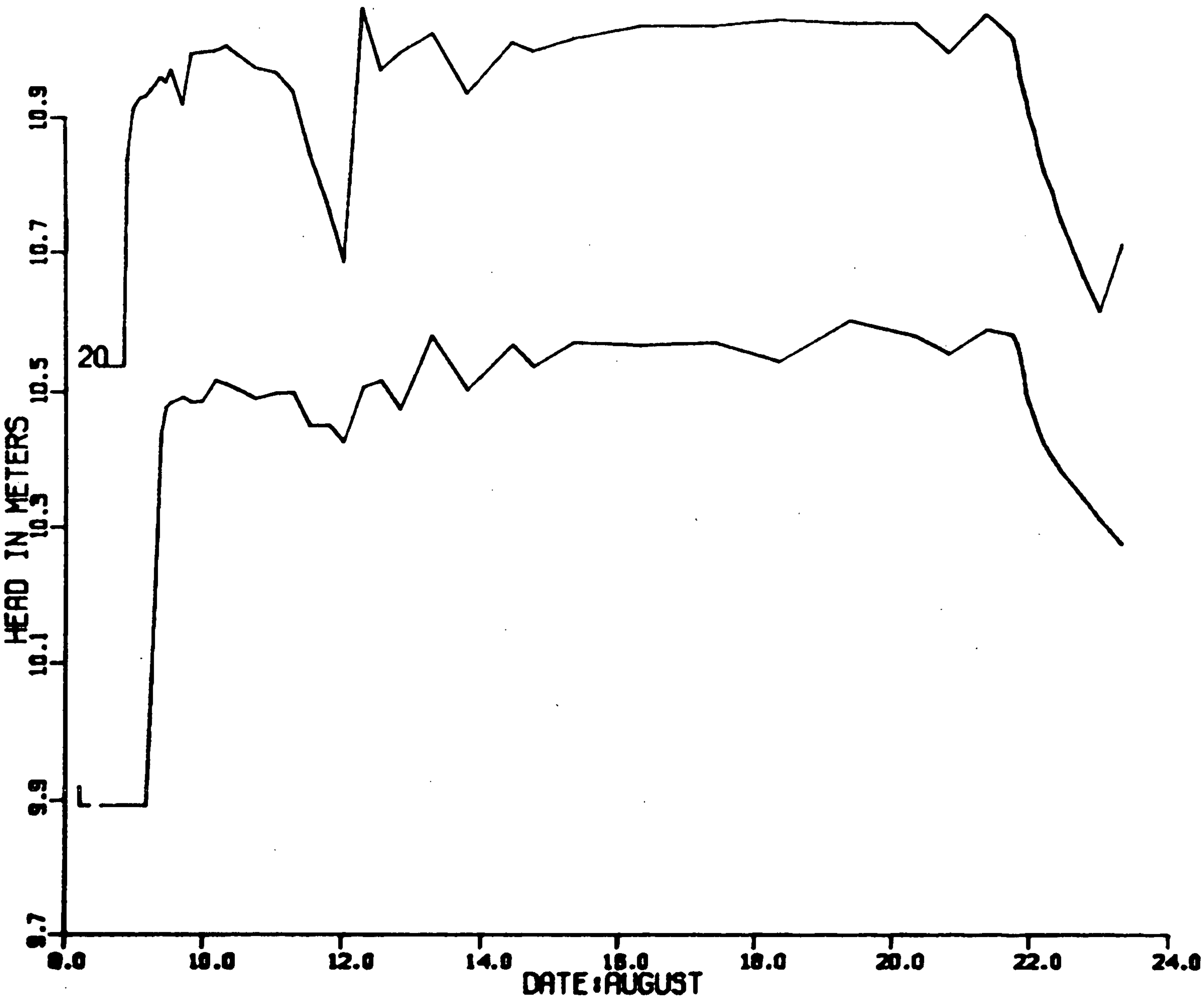

FIGURE A5-7 Hydrograph for piezometers 20 and L

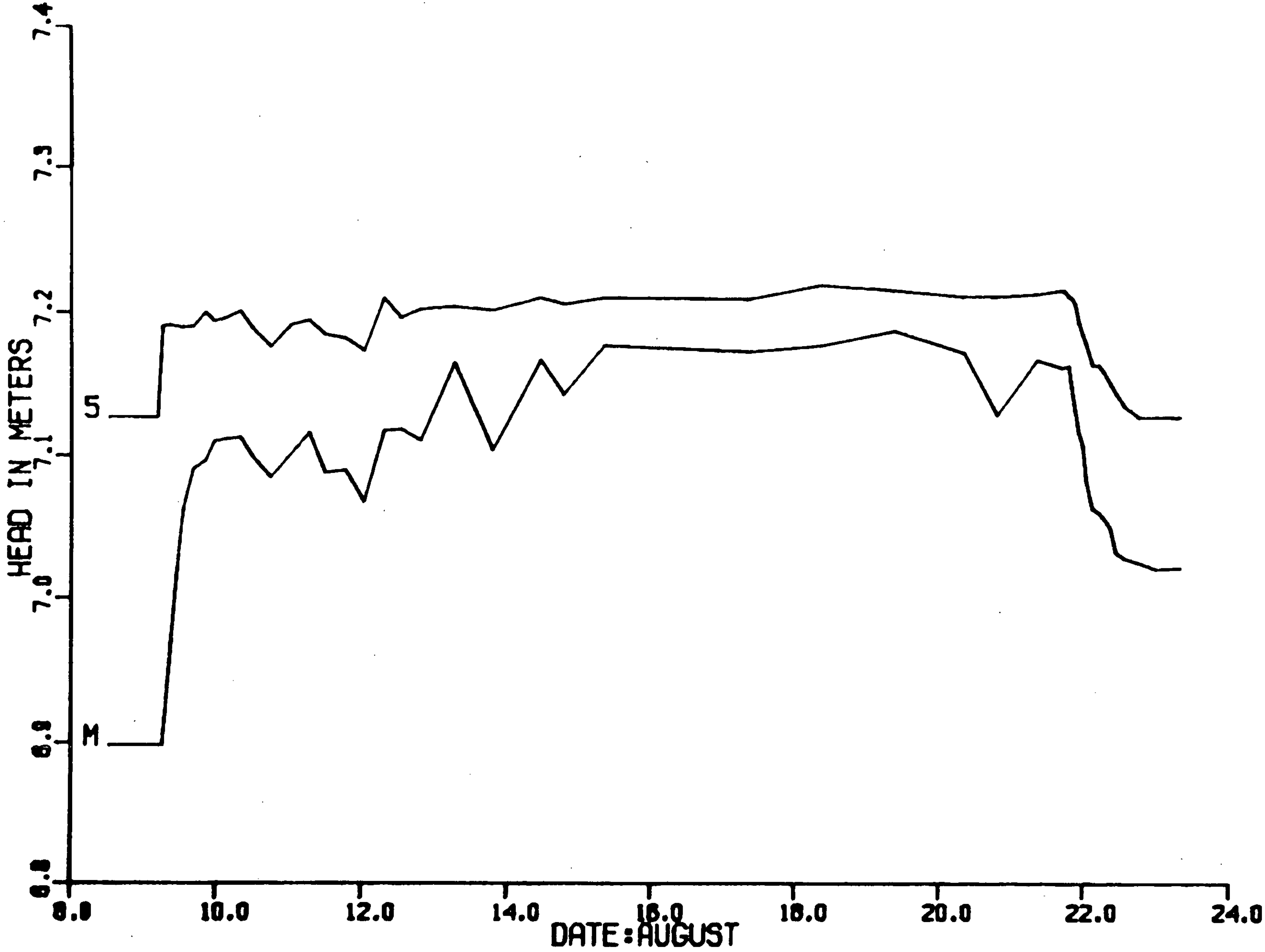

FIGURE A5-8 Hydrograph for piezometers 5 and M

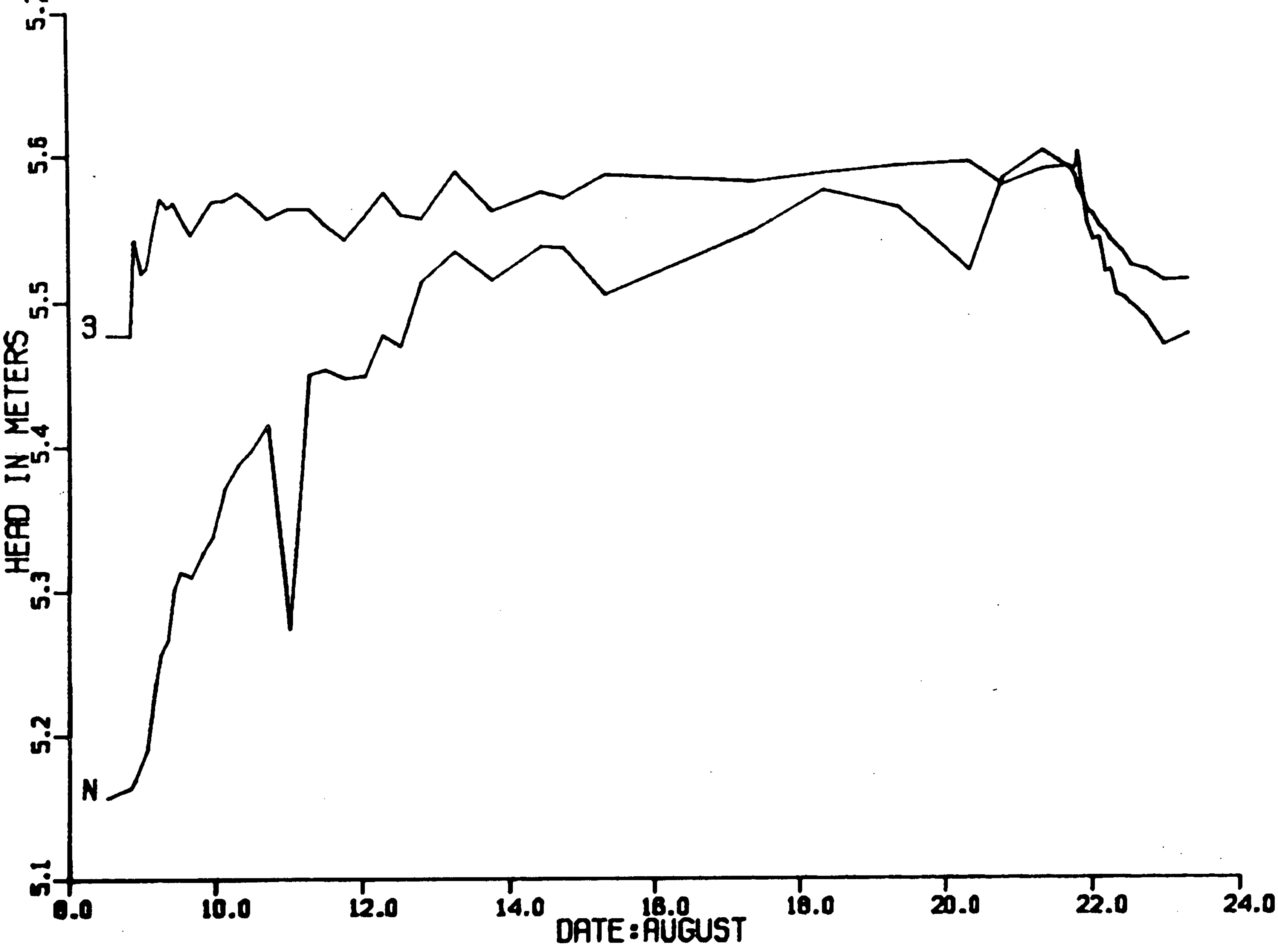

FIGURE A5-9 Hydrograph for piezometers 3 and N

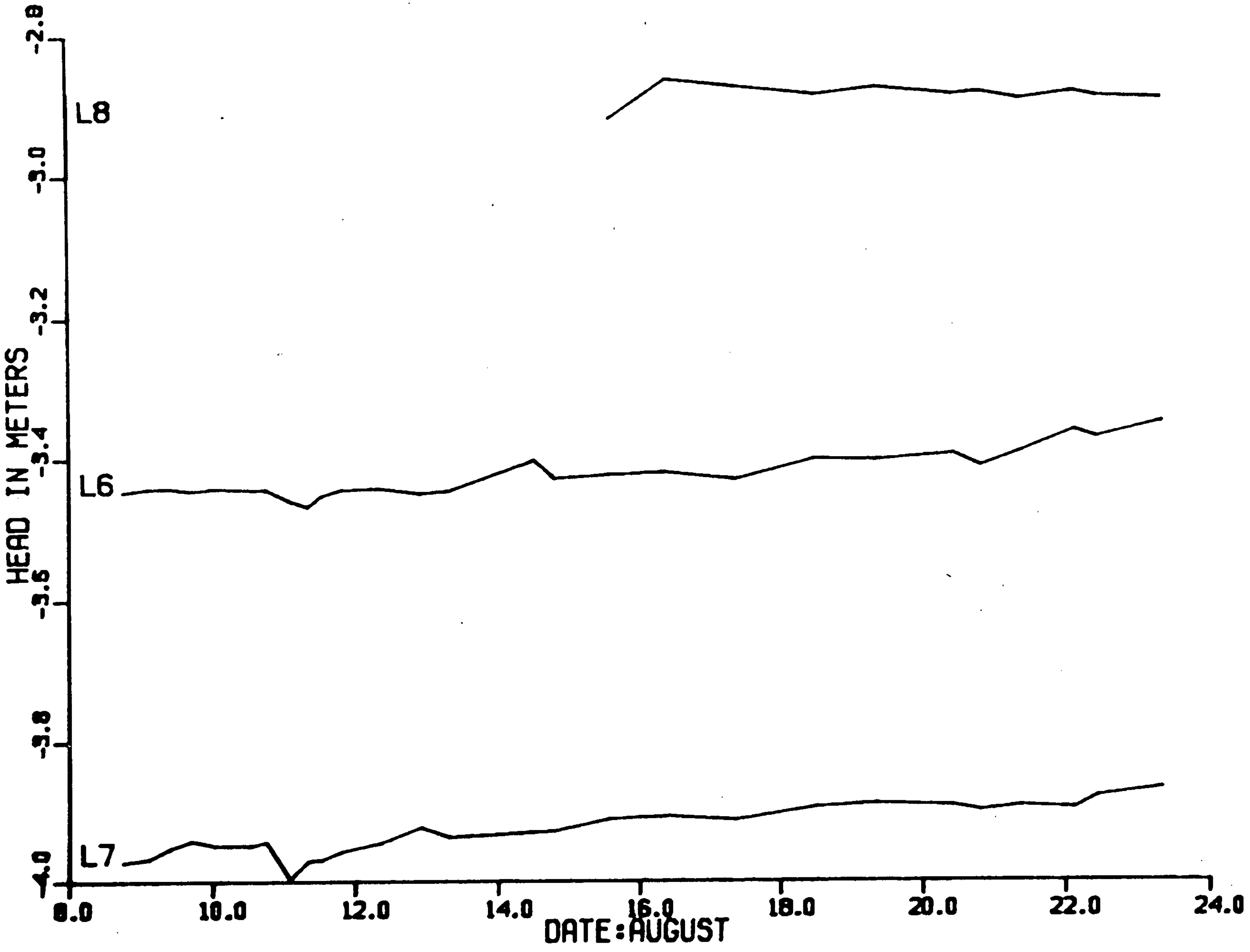

FIGURE A5-10 Hydrograph for piezometers L6, L7 and L8

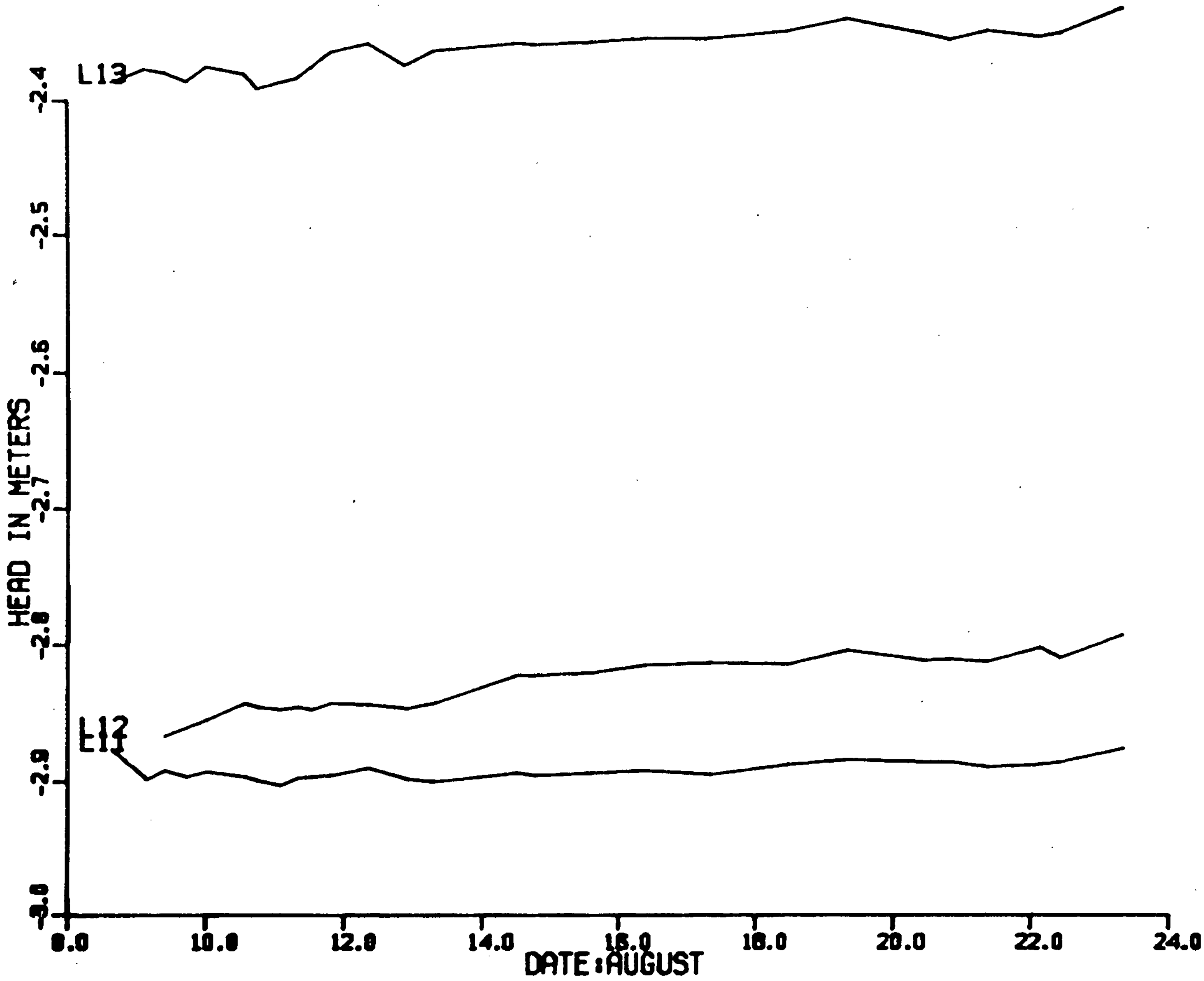

FIGURE A5-11 Hydrograph for piezometers L11,L12 and L13

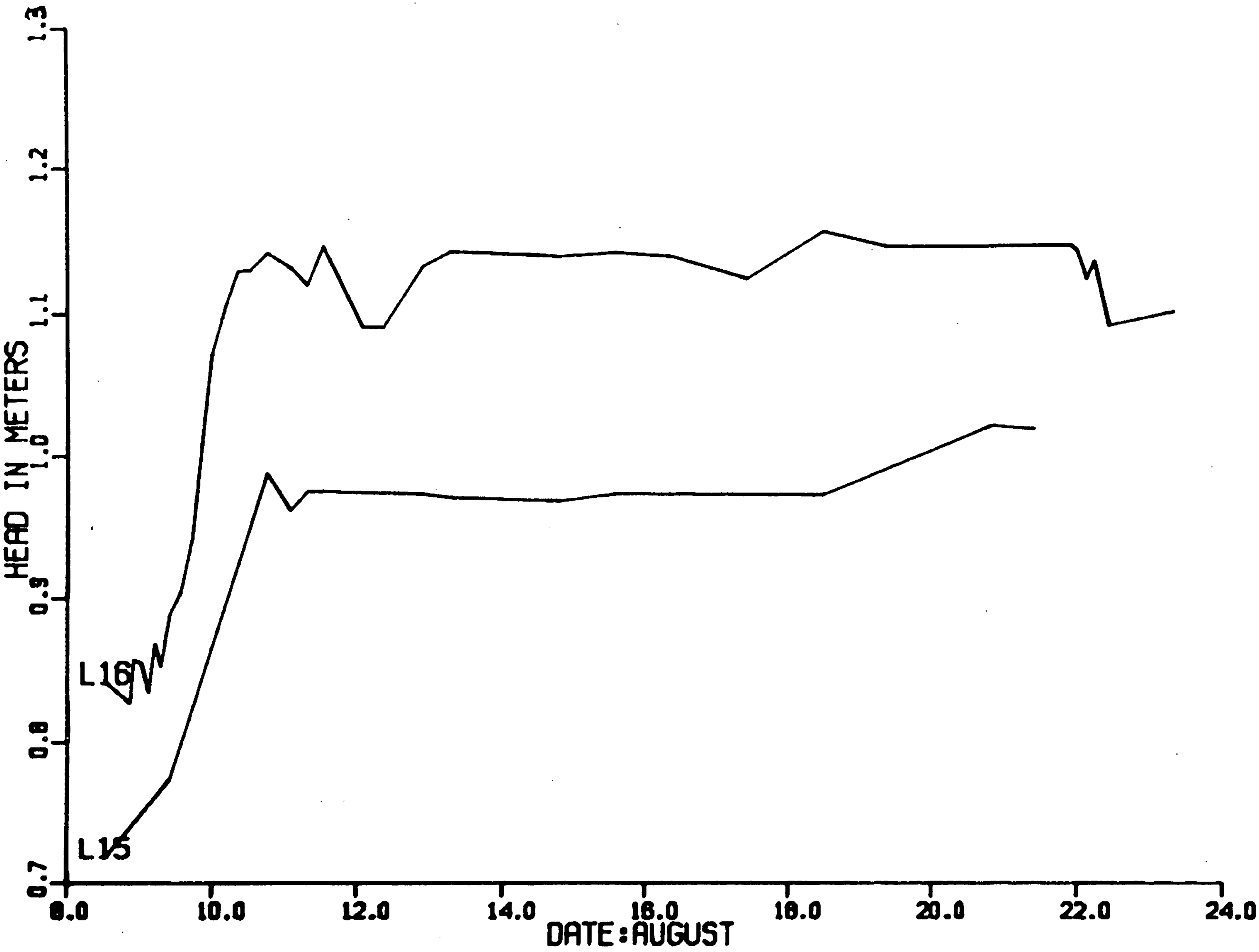

FIGURE A5-12 Hydrograph for piezometers L15 and L16

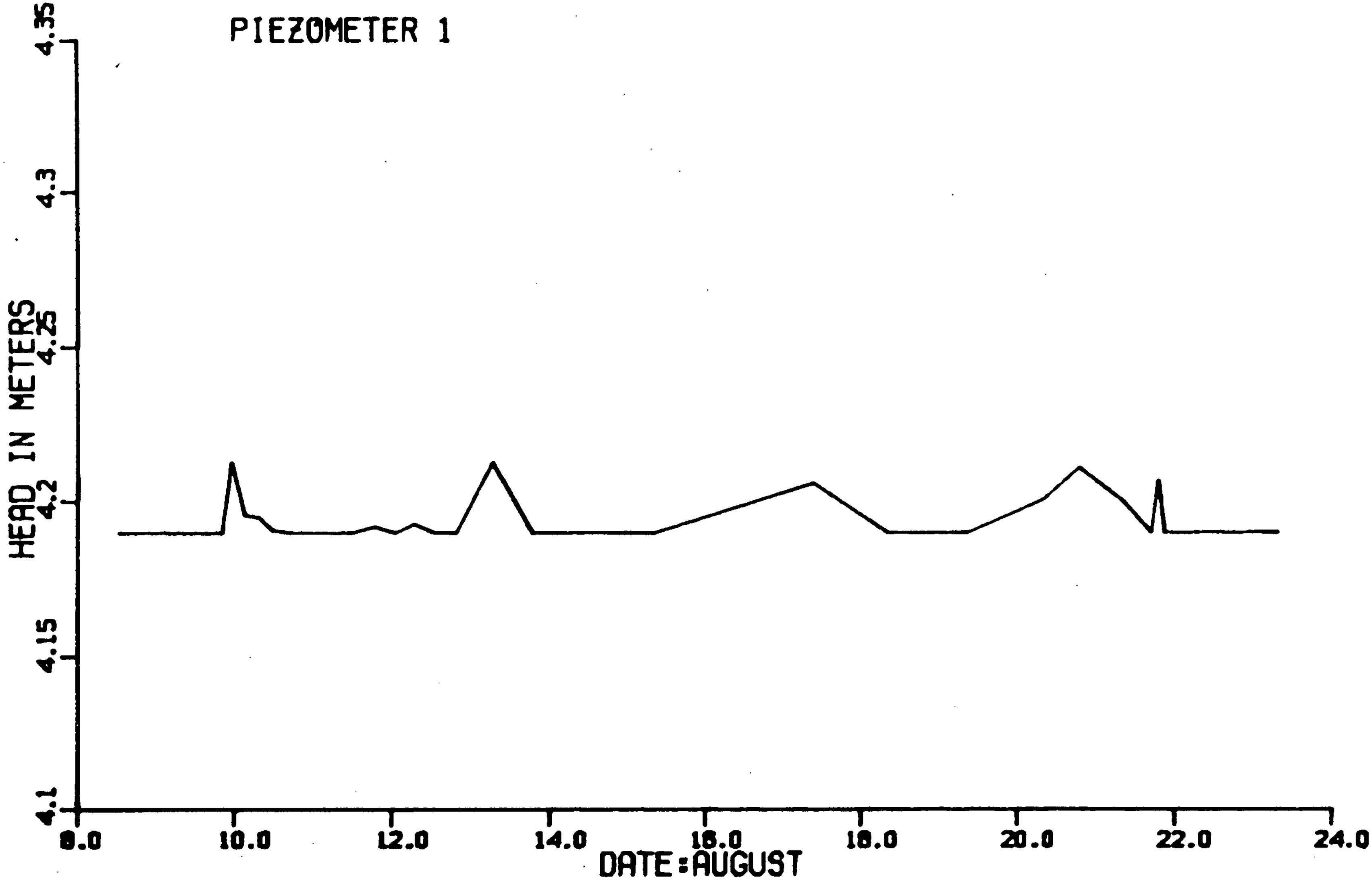

FIGURE A5-13 Hydrograph for piezometer 1

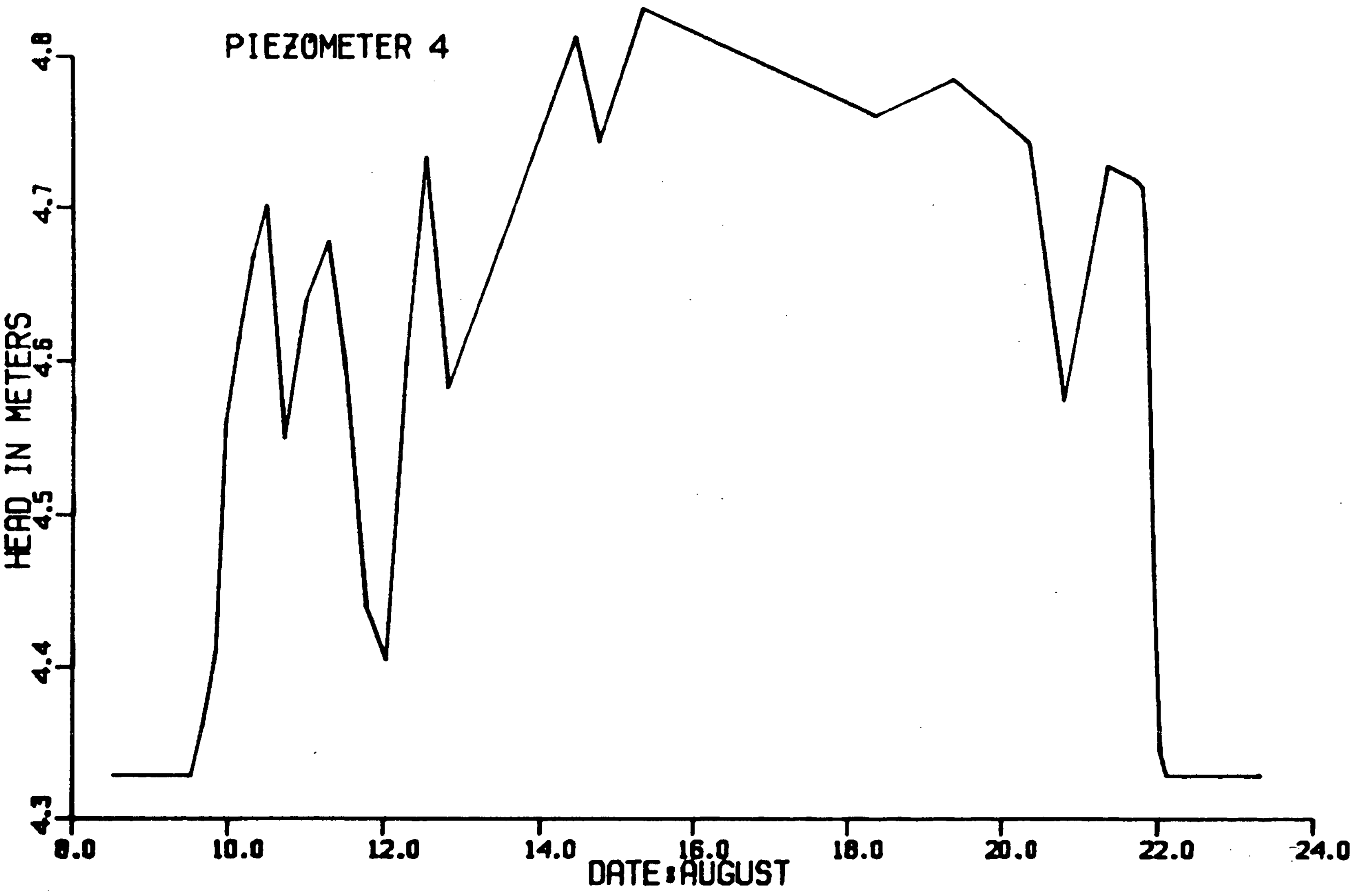

FIGURE A5-14 Hydrograph for piezometer 4

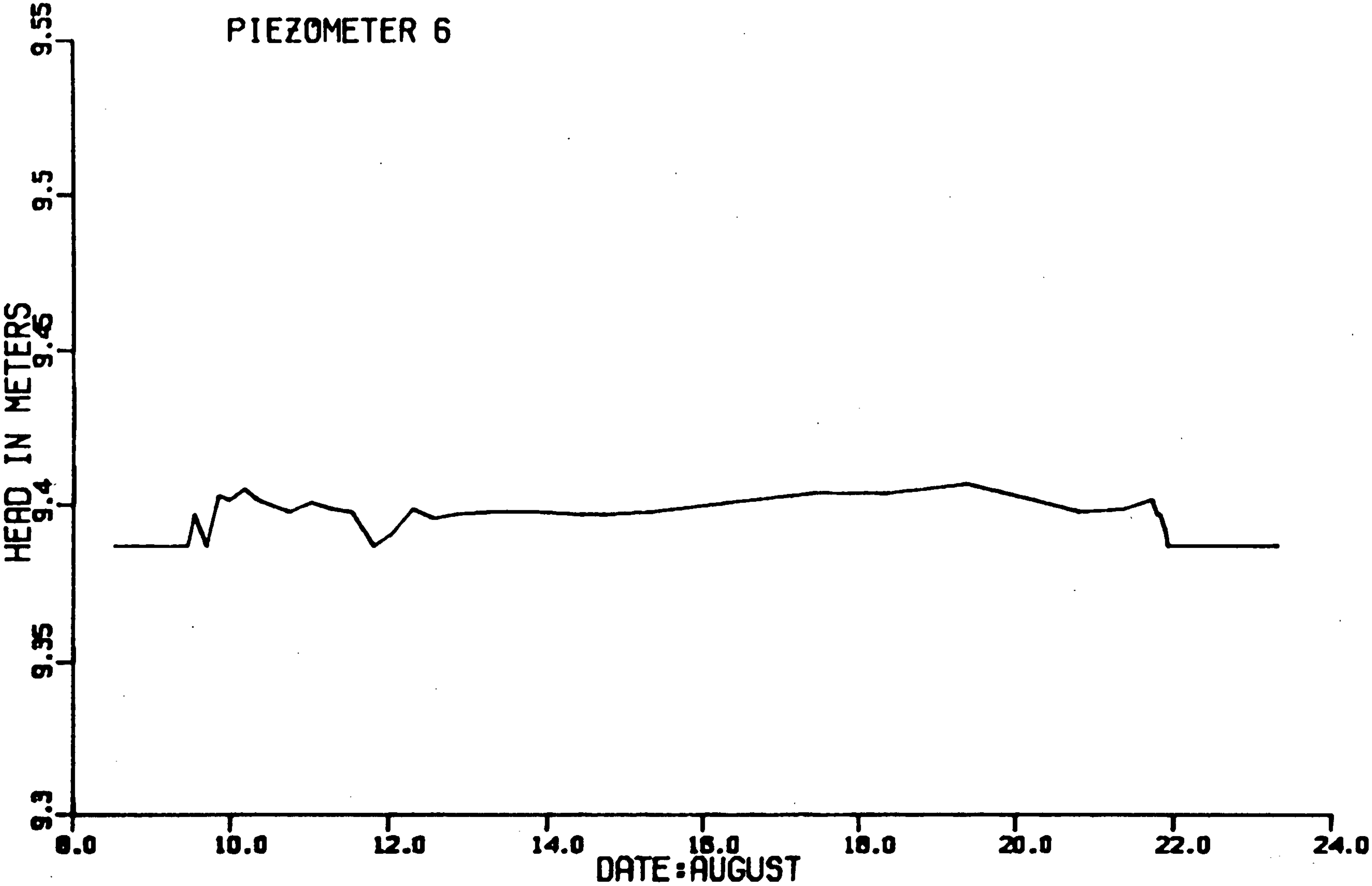

FIGURE A5-15 Hydrograph for piezometer 6

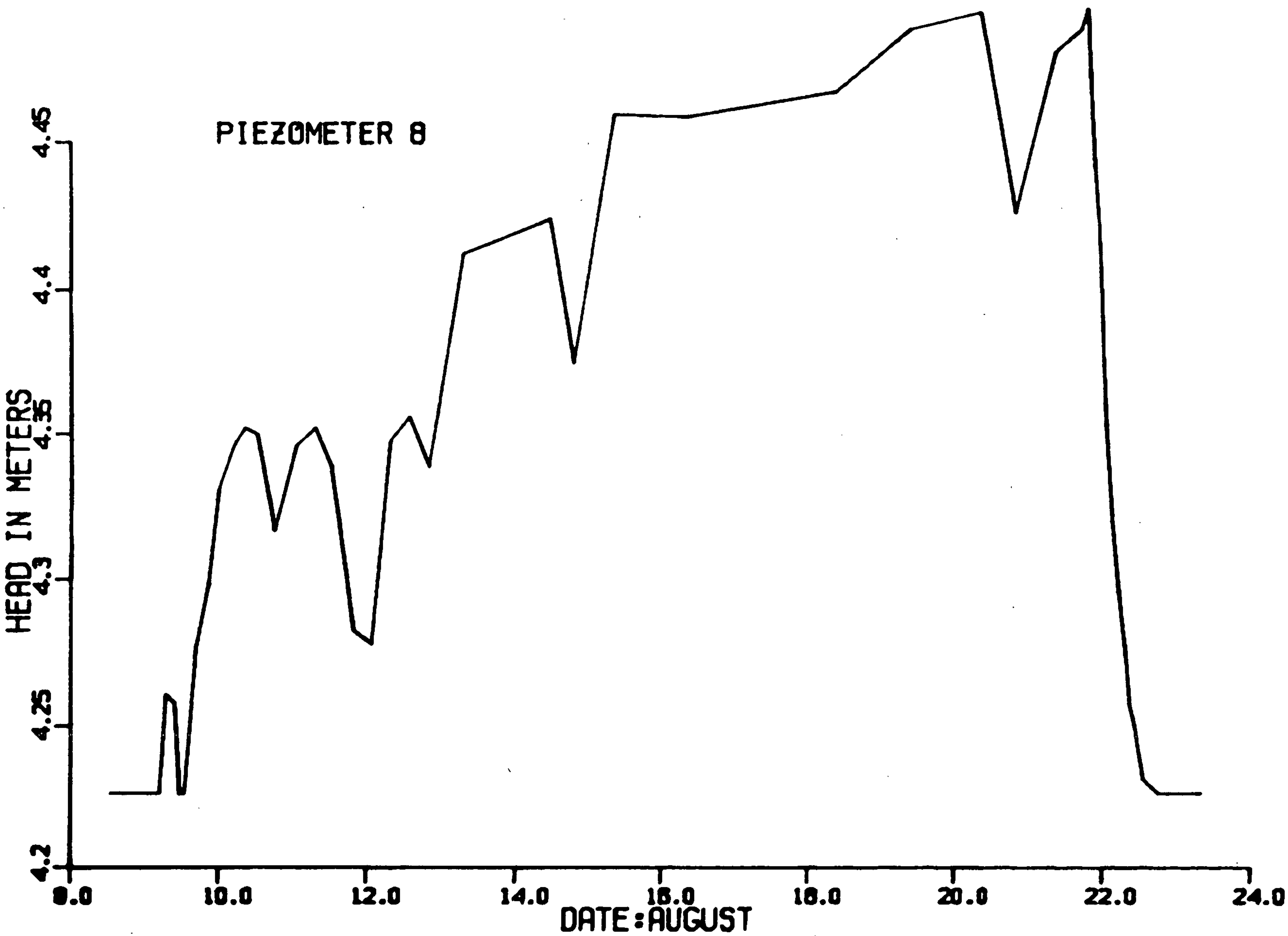

FIGURE A5-16 Hydrograph for piezometer 8

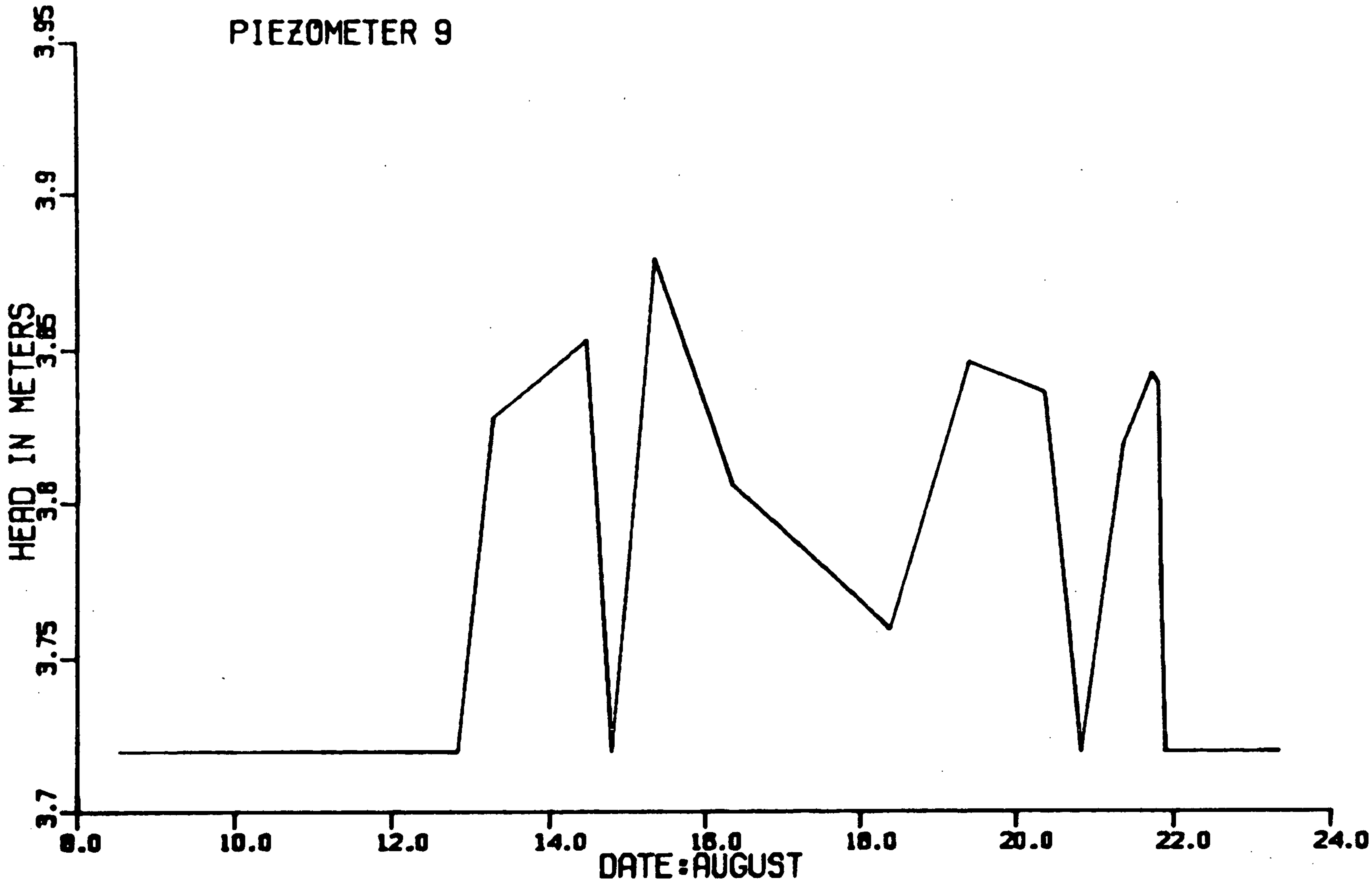

FIGURE A5-17 Hydrograph for piezometer 9

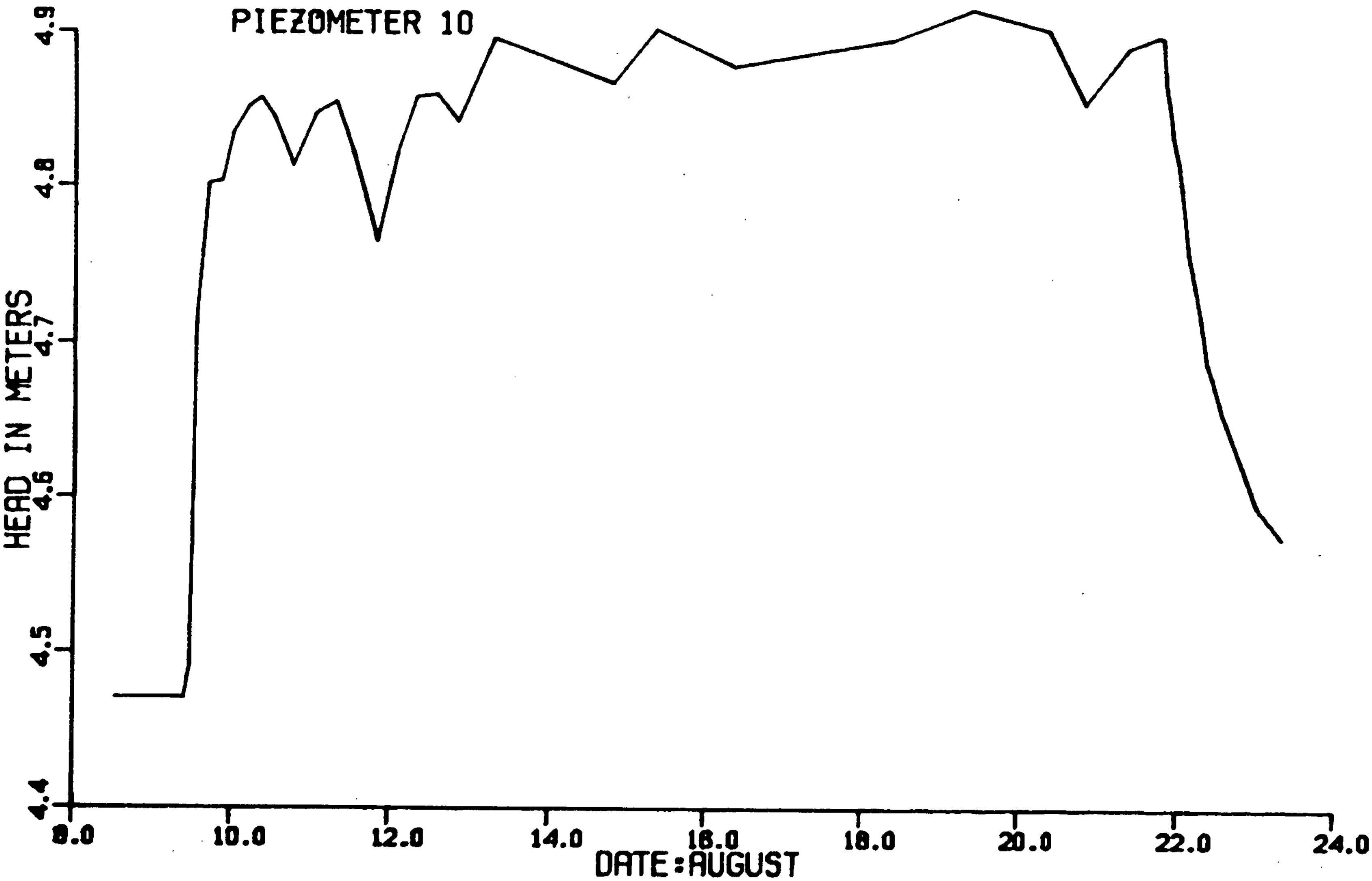

FIGURE A5-18 Hydrograph for piezometer 10

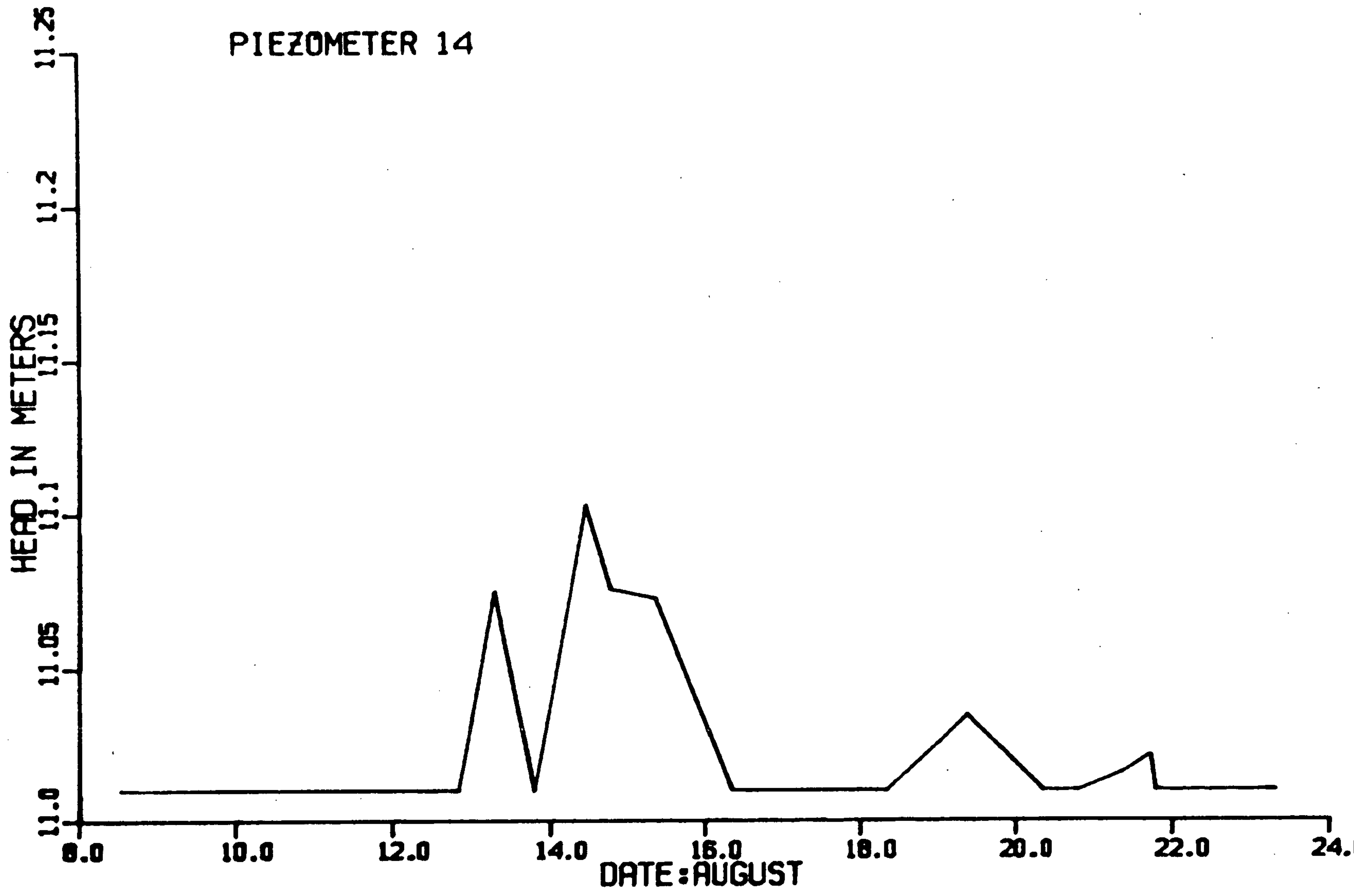

FIGURE A5-19 Hydrograph for piezometer 14

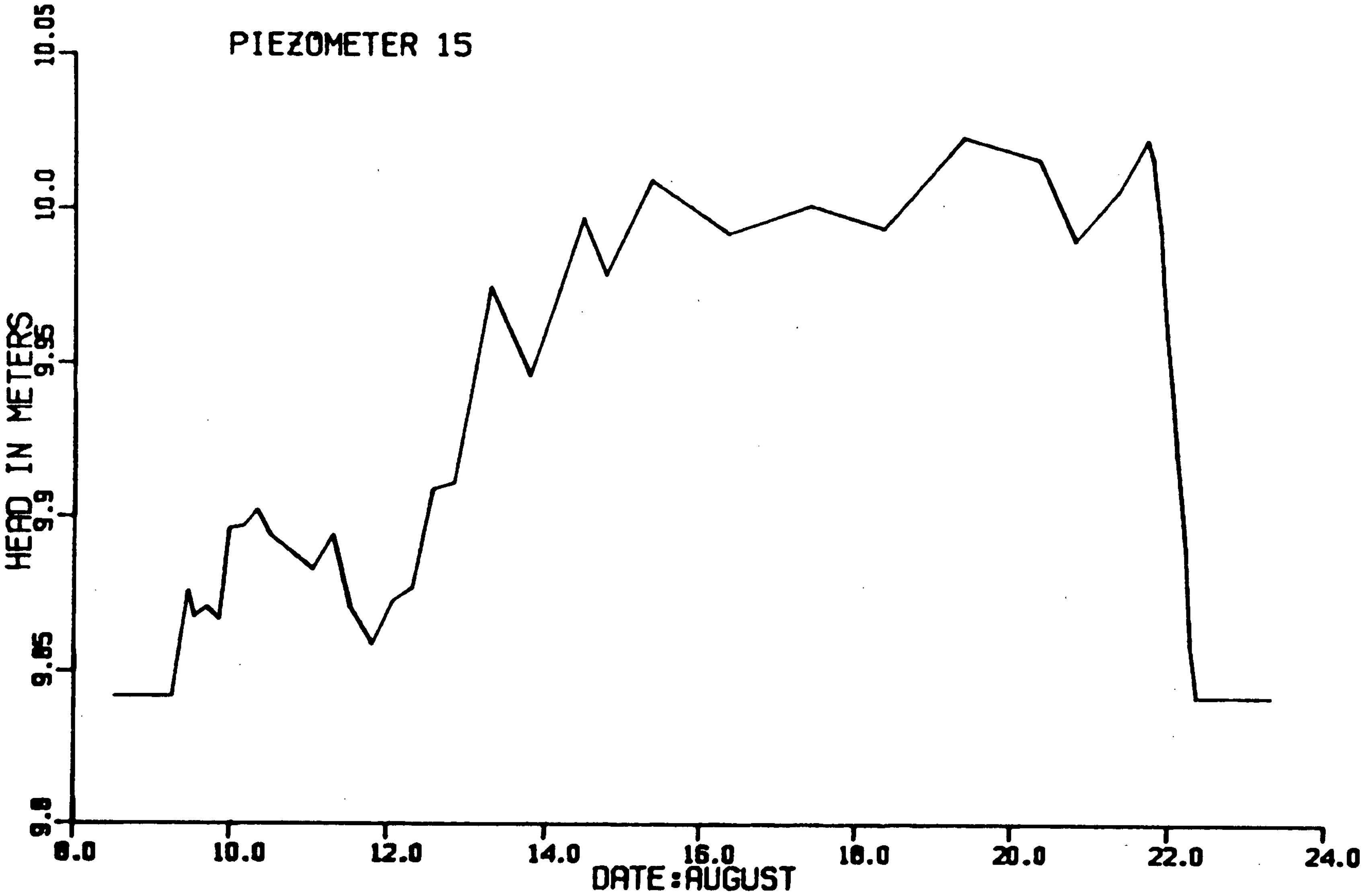

FIGURE A5-20 Hydrograph for piezometer 15

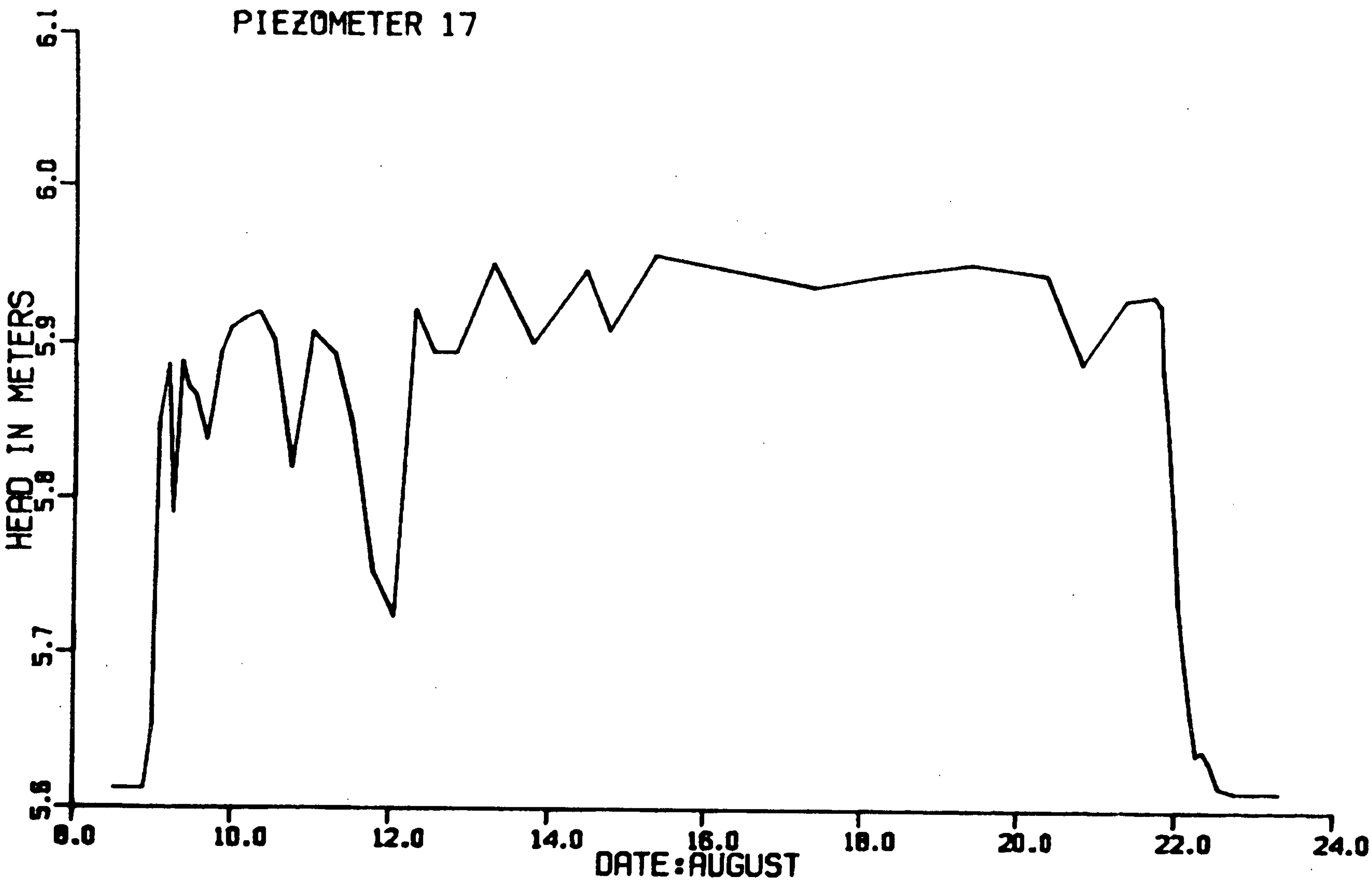

FIGURE A5-21 Hydrograph for piezometer 17

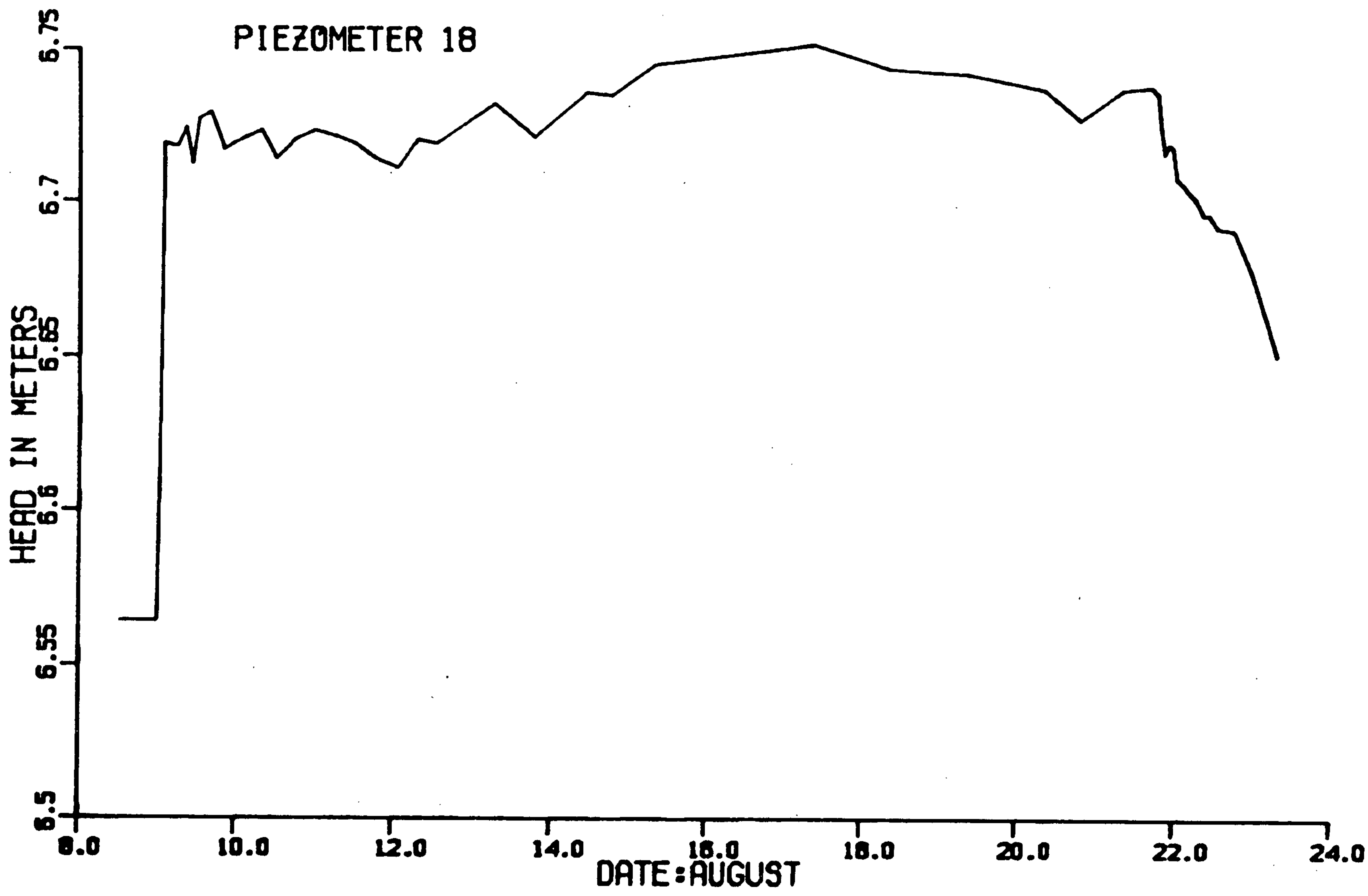

FIGURE A5-22   Hydrograph for piezometer 18

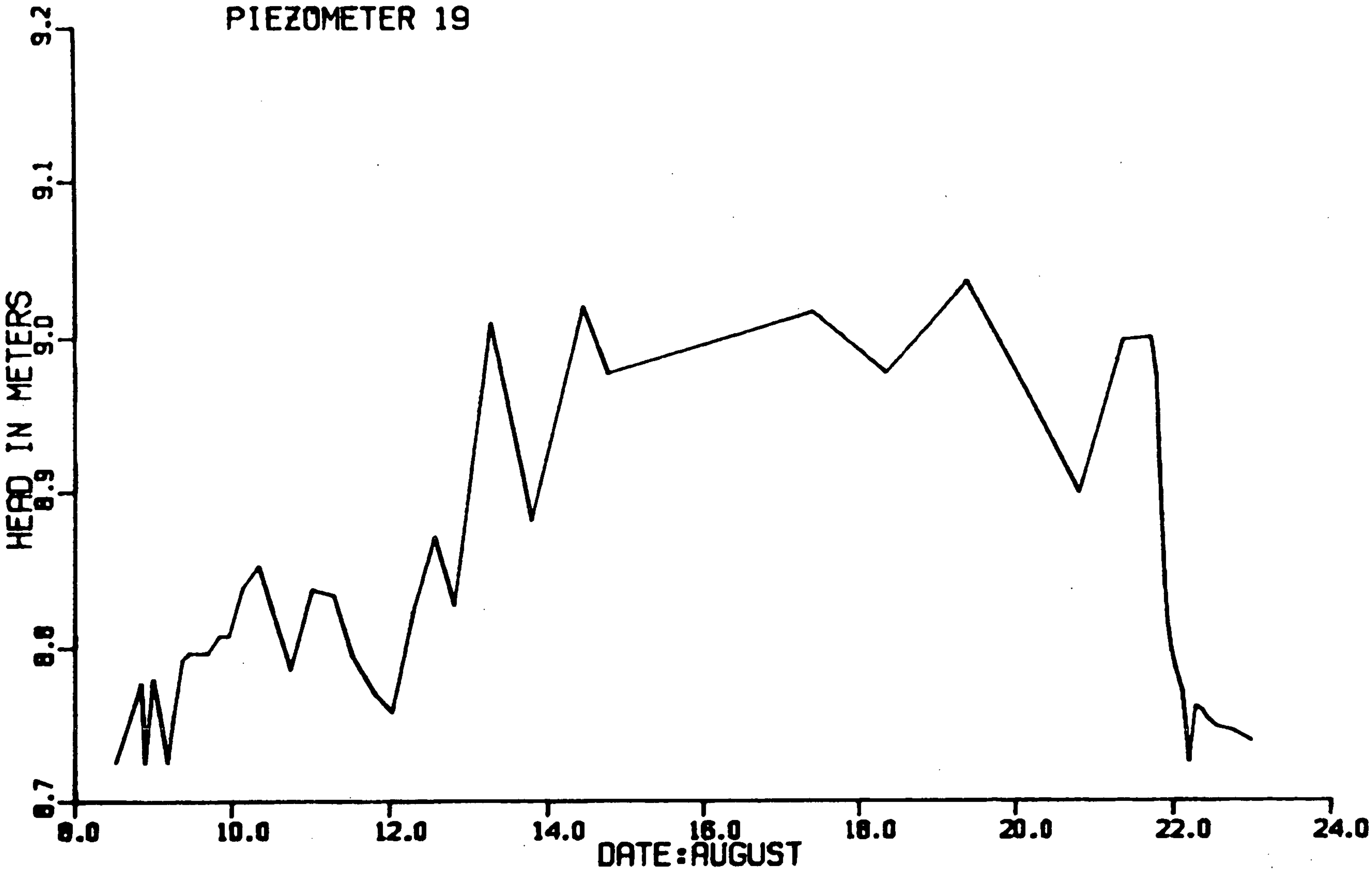

FIGURE A5-23   Hydrograph for piezometer 19