EVALUATION OF HYDRAULIC CHARACTERISTICS OF A DEEP ARTESIAN AQUIFER FROM NATURAL WATER-LEVEL FLUCTUATIONS, MIAMI, FLORIDA

by
Frederick W. Meyer
U. S. Geological Survey

Prepared by the
UNITED STATES GEOLOGICAL SURVEY
in cooperation with the
BUREAU OF GEOLOGY
FLORIDA DEPARTMENT OF NATURAL RESOURCES
and with other
CITY, COUNTY, STATE, AND FEDERAL AGENCIES

Tallahassee, Florida

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LETTER OF TRANSMITTAL

Bureau of Geology
Tallahassee
December 3, 1974

Honorable Reubin O'D. Askew, *Chairman*
Department of Natural Resources
Tallahassee, Florida

Dear Governor Askew:

We are pleased to make available the report "Evaluation of Hydrologic Characteristics of a Deep Artesian Aquifer from Natural Water-Level Fluctuations, Miami, Florida" by Frederick W. Meyer. The knowledge of the hydrologic characteristics of aquifer systems is fundamental to defining the vertical and horizontal controls on fluid movement; information which is needed for assessing the environmental impact of subsurface waste storage. This publication adds materially to this type of information and along with others to follow will provide useful background information needed to evaluate deep circulation patterns and the ultimate direction, rate of movement, and capacity of the Boulder Zone to accept injected liquid waste.

Respectfully yours,

Charles W. Hendry, Jr., *Chief*
Bureau of Geology
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August 19, 1974
Printed for the
Florida Department of Natural Resources
Division of Interior Resources
Bureau of Geology

Tallahassee
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EVALUATION OF HYDRAULIC CHARACTERISTICS OF A DEEP ARTESIAN AQUIFER FROM NATURAL WATER-LEVEL FLUCTUATIONS, MIAMI, FLORIDA

By
Frederick W. Meyer

ABSTRACT

Knowledge of the hydraulic characteristics of aquifer systems is fundamental to defining the vertical and horizontal controls on fluid movement, information which is needed for assessing the environmental impact of subsurface waste storage. To meet this objective, natural water-level fluctuations in the 2,947-foot deep Peninsula Utilities disposal well near Miami, Florida were analyzed to obtain estimates of the hydraulic diffusivity, hydraulic conductivity, specific storage, transmissivity, and the storage coefficient of the Boulder Zone. The fluctuations are caused chiefly by oceanic and earth tides, and by changes in atmospheric pressure. The oceanic tidal fluctuations probably result from loading due to tides in Biscayne Bay.

Water from the well indicates that locally the water in the Boulder Zone is chemically equivalent to sea water and has a temperature of 16°C (60.8°F). The pressure head of the salt water in the Boulder Zone at the 2,947-foot depth probably fluctuates at or near sea level. The quality and temperature of the water, and with geologic considerations, suggest that the Boulder Zone crops out in the Straits of Florida and that hydraulic connection exists between the Boulder Zone and the Straits of Florida.

The hydraulic diffusivity of the Boulder Zone is estimated to be $2.1 \times 10^{11}$ ft$^2$/day. The hydraulic conductivity and specific storage are estimated to be $2.1 \times 10^5$ ft/day and $1.0 \times 10^{-6}$ per foot, respectively, based on an assumed porosity of 50 percent and a 15-foot thickness of aquifer. The transmissivity and storage coefficient are estimated to be $3.2 \times 10^6$ ft$^2$/day and $1.5 \times 10^{-5}$, respectively.

INTRODUCTION

As part of man's attempt to solve the increasing problem of liquid-waste disposal and to alleviate the environmental deterioration of fresh and estuarine waters, deep-well injection is being considered and evaluated. Southern Florida is underlain at great depths by highly permeable carbonate aquifers and interbedded fine-grained, less permeable strata. These hydrologic conditions may be favorable to the successful use of deep disposal wells.
Extensive cavern systems, some of which are indicated by zones of lost circulation of drilling fluids, have been reported throughout southern Florida. Zones of highly permeable dolomite and limestone collectively called the Boulder Zone (Kohout, 1965 p. 256), contain relatively cold saline water at depths of about 3,000 feet at the coast near Miami. The occurrence of abnormally cold sea water at depth and high transmissivities within the Boulder Zone suggest hydraulic connection with the Straits of Florida, 30 miles east of Miami.

During 1969, a 2,947-foot disposal well was drilled near Miami at the Peninsula Utilities (a subsidiary of General Waterworks Corporation) Snapper Creek Park sewage-treatment plant. Currently (1973), secondarily treated effluent is injected down the well into cavernous limestone at a rate of 2 million gallons per day.

Although data are available on the subsurface geology, water quality, and injection rates (Garcia-Bengochea, 1970; Vernon, 1970), no analyses have been made to determine the hydraulic characteristics of the Boulder Zone. To overcome this deficiency natural water-level fluctuations in the disposal well were analyzed to obtain preliminary estimates of the Boulder Zone's transmissivity and storage coefficient, and to determine the degree of hydraulic separation with the overlying brackish and fresh-water zones and the degree of hydraulic connection of the Boulder Zone with the Straits of Florida. This will provide useful background information needed to evaluate deep circulation patterns and the ultimate direction, rate of movement, and capacity of the Boulder Zone to accept injected liquid waste. This preliminary analysis does not eliminate the need for controlled pumping tests and additional exploratory drilling.

PURPOSE AND SCOPE

The U. S. Geological Survey, as part of its research program to evaluate the effects of underground waste disposal on the Nation's subsurface environment, began the collection and analysis of water-level data from the Peninsula well during 1970 and 1971.

The purposes of this investigation were (1) to determine the pressure head of salt water in the Boulder Zone at the Peninsula well, (2) to identify the causes and determine the magnitudes of the several types of water-level fluctuations in the well, and (3) to determine the transmissivity and storage coefficient of the Boulder Zone by an analysis of natural water-level fluctuations.
A continuous record of water-level fluctuation in the Boulder Zone was obtained at the Peninsula well during February 12 to March 1, 1970; and from November 25, 1970 to February 12, 1971. Concurrent data on the tides in the ocean at Miami Beach and in Biscayne Bay at Coconut Grove were obtained from stage stations operated by the National Ocean Survey and the U. S. Geological Survey, respectively. Personnel of the Peninsula Utilities Company determined the altitude of the land surface at the well. Barometric data were obtained from the U. S. Weather Service at Miami International Airport, 7 miles northeast of the well site.

Measurements of head in the principal artesian water-bearing zones that overlie the Boulder Zone were obtained on February 10 and 11, 1971 by use of a mercury manometer on the outlet to the annular space between the casings in the well. Details of the well construction were described by Jose I. Garcia-Bengochea in an engineering report to General Waterworks Corporation (1970).

The data were analyzed using methods described by Ferris (1951), Carr and Van Der Kamp (1969), and Van Der Kamp (1972). Hilton H. Cooper and E. P. Weeks, Geological Survey research scientists, suggested a method to separate earth and oceanic tidal components of the observed semidiurnal water-level fluctuations in the well, and a method to calculate hydraulic diffusivity (written commun., 1973).

ACKNOWLEDGMENTS

The author expresses his appreciation to Clarence L. Reynolds, Vice-President of the Southern Division of General Waterworks Corporation, Coral Gables, Florida, and Jose I. Garcia-Bengochea, Vice President of Black, Crow, and Eidsness Consulting Engineers, Gainesville, Florida, for their cooperation in obtaining the necessary data, and to Hilton H. Cooper, Edwin P. Weeks, Charles A. Appel, John E. Hull, and Howard Klein, colleagues in the U. S. Geological Survey for assistance in collecting and analyzing the data.

This investigation was supported as part of the U. S. Geological Survey's nationwide subsurface waste storage research program and as part of the statewide cooperative water resources program.
LOCATION AND GEOFHYDROLOGIC SETTING

The Peninsula well is in Dade County, about 10 miles southwest of Miami (fig. 1). It is 2,927 feet deep and is cased to 1,810 feet (fig. 2). The land surface at the well is about 6 feet above msl (National Ocean Survey, mean sea-level datum 1929).

The local water supply is obtained from the Biscayne aquifer, a highly permeable limestone strata that underlies the area to a depth of about 100 feet. Beneath the Biscayne aquifer is a 300-foot thick confining bed composed of sand and clay, which confines the water in the underlying Floridan aquifer system. The Floridan is about 1,500 feet thick and is composed of several

Figure 1  Map showing site location.
hydraulically separate water-bearing zones (Meyer, 1971). The upper 600-foot section is composed of limestone interbedded with calcareous clay and the lower 900-foot section (the principal water-bearing zone) is composed chiefly of highly permeable dolomitic limestone. The head and the salinity of the ground water increase with depth in the Floridan aquifer. Locally the head of the brackish water in the principal artesian water-bearing zone stands 41 feet above msl.

Figure 2 Sketch showing well construction.
Chloride concentrations in the water from this zone range from 1,000 to 4,000 mg/l (milligrams per liter).

A 1,000-foot thick confining bed composed of limestone interbedded with dolomite underlies the principal artesian water-bearing zone. The permeability of the bed varies with depth but generally is low in the vertical direction, according to Vernon (1970, p. 10). Chloride concentrations of water in the confining bed range from 4,000 to 18,000 mg/l; and the zone of transition between brackish and saline water (sea-water) occurs between 1,800 and 2,000 feet below sea level (Garcia-Bengochea, 1970).

The Boulder Zone, consisting of cavernous strata, is at the bottom of the Peninsula well. The Bureau of Geology, Florida Department of Natural Resources, designated the Boulder Zone as the most favorable place to inject treated liquid wastes (Vernon, 1970, p. 31). According to the Department of Natural Resources (Vernon, 1970, p. 23) the requisites for deep disposal are "...that the wells terminate in zones of high transmissivities that are filled by saline or unusable water. Such zones are located in the base or below the Floridan aquifer and are reasonably separated from usable waters by dense sediment with low or minimal vertical transmissivities". The terms "saline water" and "unusable water" need to be defined. Water containing more than 1,000 mg/l of dissolved solids is generally considered to be "saline" by the U.S. Geological Survey. However, a criteria based on 15,000 mg/l might be more appropriate if the saline water can be economically used, that is, economically converted to fresh water.

An extension of the subsurface geology from the Everglades eastward to Bimini, in the Bahamas, suggests that the Boulder Zone crops out in the Straits of Florida some 35 miles east of the well (fig. 3). Dredge hauls and seismic profiles indicate the principal artesian water-bearing zone in the Floridan aquifer system also crop out in the Straits of Florida (Malloy and Hurley, 1970, p. 1970). Malloy and Hurley (1970, p. 1947) indicate that karst features (sinkholes) found on the east slope of the Straits may be kept free of sediment by submarine flow from the Floridan aquifer.

Analysis of a water sample from the bottom of the well indicates that locally the water in the Boulder Zone is chemically equivalent to sea water. Temperature data obtained during drilling (Garcia-Bengochea, 1970, plate 4-2) indicate that ground water becomes cooler with depth rather than warmer. The thermal gradient in ground water seems to be closely related to the thermal gradient in the ocean in the adjacent Straits of Florida (fig. 3).
A hypothesis that geothermal heating induces a cyclic flow of cold dense sea water from the Straits of Florida inland through cavernous dolomitic limestone was suggested by Kohout (1965 and 1967). Also, an inland flow of sea water by dispersion-induced circulation may be indicated, as suggested by Cooper and others (1964). In both cases such flow would require that the head in the well in terms of an equivalent column of sea water (hence length and density) would be lower than the sea.

Alternatively, Vernon (1970) and Garcia-Bengochea (1970) have suggested that the temperature-gradient anomaly reflects the loss of heat from the aquifer by conduction into the cold deep water of the Straits of Florida. Vernon (1970, p. 13) states "Since no head exists at that depth, no flow and exchange of water
with the ocean would be expected. Cooler temperatures are to be expected in zones of large solutional caverns where larger volumes of water and increased velocities of flow are present, when compared to zones of lower transmissivities.”

To date (1973), the direction and amount of flow, if any, can only be speculated upon because of the lack of definitive data on hydraulic gradients and the horizontal extent and interconnection of the caverns. No pumping tests have been made to determine the hydraulic characteristics of the Boulder Zone. Therefore an analysis of natural water-level fluctuations may provide useful information on the hydraulic characteristics, including transmissivity of the aquifer system and the extent of hydraulic connection with the adjacent Straits of Florida.

WATER-LEVEL FLUCTUATIONS IN THE BOULDER ZONE

Fluctuations of the water level in the Peninsula well were continuously recorded from February 12 to March 1, 1970, and from November 25, 1970 to February 12, 1971. Long-term water-level fluctuations, lasting weeks or months, were related chiefly to changes in the density of the water column in the well due to dispersion of salt water and thermal conduction, to changes in atmospheric (barometric) pressure, and, perhaps, to seasonal changes in tides. Short-term cyclic fluctuations, lasting hours or days were related chiefly to ocean and earth tides, and to atmospheric tide.

A summary of events before and after collection of water-level data follows to help explain some of the variation in the water levels. Construction of the Peninsula well began June 19, 1969, and ended December 15, 1969. The well was drilled to a depth of 2,947 feet and contained 1,810 feet of 16-inch steel casing. On December 16, 1969, the well was pumped at about 3,000 gpm (gallons per minute) with compressed air to clear cuttings from the hole, and caliper and fluid-velocity logs were obtained by the Florida Bureau of Geology. On December 17, 1969, the well was pumped for about 2 hours at 3,000 gpm and a water sample was collected for analysis by the U. S. Geological Survey. The temperature of the sample was 16°C (60.8°F), and its chloride content was 19,300 mg/l. The water level in the well was 0.4 foot above msl 35 minutes after pumping ceased. Subsequent measurements indicated a slow rise in the water level.

From January 9 through January 29, 1970, fresh water from a nearby canal was intermittently injected into the well at rates ranging from 1,000 to 4,000 gpm for testing. On February 6, 1970, cement bond, electric, and gamma-ray logs were obtained. During February 12 through March 1, 1970,
records of water-level fluctuations were obtained by the U. S. Geological Survey. On March 25, 1970, a repeat caliper log was obtained by the Florida Bureau of Geology. On April 23, 1970, fluid conductivity, gamma ray, electric, and temperature logs were obtained by the U. S. Geological Survey. During November 25, 1970, through February 12, 1971, records of water levels were obtained by the U. S. Geological Survey. On January 21, 1971, temperature and casing-locator logs were obtained by the Florida Bureau of Geology. During March through June 1971, the monitoring and injection systems were connected to the well, and effluent from the treatment plant was intermittently injected into the well for testing. On July 23, 1971, the system became operational, and effluent flows of 2 to 3 mgd have been successfully injected into the well.

**LONG-TERM FLUCTUATIONS**

The daily mean water level ranged from 6.8 to 7.1 feet above msl from February 12 to March 1, 1970; from 3.4 to 3.9 feet above msl from November 25, 1970 to January 20, 1971; and between 5.6 and 5.9 feet above msl from January 21 to February 12, 1971 (fig. 4).

In December 1969 the water level in the well was about 0.4 foot above msl after the well had been pumped to obtain a water sample. An analysis of the sample indicated that it was similar to sea water (chloride concentration of 19,300 mg/l).

In January 1970, during an injection test, a large amount of warm fresh canal water was pumped into the well. After the test, the injected fresh water was allowed to backflow into the canal. Flow ceased when the salinity of the water column approached that of the Boulder Zone water. The water level in the well from February 12 to March 1, 1970 was about 7 feet above msl (fig. 4).

However, on March 18, 1970, the chloride content of the water in the upper 100 feet of the well was 18,000 mg/l, or about 1,300 mg/l less that in the Boulder Zone. A fluid-conductivity log obtained on April 23, 1970, indicated that the salinity increased gradually with depth in the water column in the well. Therefore the water level, in terms of the density of the Boulder Zone water, was probably as much as 6 feet below the observed water level of February 12 to March 1, 1970, or about 1 foot above msl. This compares with 0.4 foot above msl measured December 17, 1969, when the water in the well was about the same density as the water in the Boulder Zone due to recent pumping.

From March 1 to November 25, 1970, when daily records were not available, the water level declined 3 feet, which, if the decline were steady,
Figure 4 Hydrograph of water-level fluctuations in the Peninsula well, February 12 - March 1, 1970; and November 25, 1970 - February 12, 1971.
would have been at a rate of 0.012 foot per day. The decline was probably due chiefly to the dispersion of salt water from the Boulder Zone into the upper part of the water column in the well and to cooling of the water column. Hence, the decline was caused by a gradual increase in the density of the water column.

From November 25, 1970, to January 20, 1971, the water level declined from 3.9 to 3.4 feet, or an average rate of 0.008 foot per day. The recession was probably related to the gradual increase in density, and in part to the seasonal decline in sea level, as shown by the comparison with the ocean and bay hydrographs in figure 5.

On January 21, 1971, after an electric log of the well was obtained by the Florida Bureau of Geology, an unknown amount of relatively warm fresh water was allowed to flow into the well while the logging cable was being rinsed. Upon completion of the rinsing, the water level in the well stood at 5.7 ft above msl (fig. 4) due to the addition of the less dense water to the top of the water column.

The wide variation in water levels during these periods was apparently caused by operationally induced changes in the density of the water column in the well. Although declining, the water level was fluctuating naturally in response to changes in atmospheric pressure and to tides (figs. 5 - 8).

**ALTITUDE OF THE WATER LEVEL**

Wide variations in the altitude of the water level in the well were indicated by water-level measurements made on December 17, 1969 before injection of fresh water into the well, and by continuous records of water-level fluctuations during February 12 through March 1, 1970, and November 25, 1970, through February 12, 1971, after injection of fresh water into the well (figs. 4 and 5). On December 17, 1969, the water level was observed to be rising slowly from an altitude of 0.4 foot above msl only 35 minutes after the well had been pumped at about 3,000 gpm. The rise in water level could be related to the recovery of the water level in the well after pumping, to changes in the density of the water column, or to a combination of both. The cavernous nature of the aquifer suggests that the water level would recover quickly; therefore, the slow rise in the water level is, in the author's opinion, an indication of decreasing density in the water column. The decrease in density could be caused by warming of the water column in the well or by inflow of less saline water into the water column from strata exposed to the well between the bottom of the casing at 1,810 feet and the cavern at 2,930 feet, or to a combination of both causes.
Therefore, the variations in the altitude of the water level in the well, as shown in Figure 4, did not represent natural changes in pressure head of the salt water in the Boulder Zone. Variations in the salinity of the water column indicate that the water level, in terms of a well filled with water of the same density as that in the Boulder Zone, probably fluctuates naturally at or near present mean sea level.

Accurate pressure-head determinations are extremely important for evaluating hydraulic gradients and deep circulation patterns. The true pressure head can be measured by (1) installing a 2,930-foot liner inside the well with a packer, (2) pumping the well until the temperature and salinity of the water column in the liner are the same as that of the Boulder Zone water, (3) observing the short-term recovery from pumping until the water level becomes more or less static, and (4) correcting for changes in temperature. Also, the pressure head can be accurately measured by installing a sensitive pressure transducer at the bottom of the well.

ATMOSPHERIC EFFECTS

A comparison of the daily mean water level in the well with the daily mean barometric pressure (expressed in feet of water) at Miami (fig. 5) indicates that water-level fluctuations ranging from a day to several days duration were chiefly caused by changes in atmospheric pressure. A comparison of barometric fluctuations with sea-level fluctuations (fig. 5) indicates that the ocean-level also fluctuated in response to changes in atmospheric pressure. However, the atmospheric effect on the water level in the Peninsula well is shown more clearly in Figure 6. The graphs indicate that decreasing atmospheric pressure caused the water level in the well to rise and that increasing atmospheric pressure caused the water level to decline.

The barometric efficiency (BE) of the Boulder Zone was estimated, by the author, to be 0.70 by using the equation (Ferris and others, 1962, p. 85)

\[
BE = \frac{S_w}{S_b}
\]  

where \( S_w \) is the net change in the water level in the Peninsula well from February 19 to 21, 1970, and \( S_b \) is the corresponding change in atmospheric pressure, both expressed in feet of water.
Figure 6  Graph comparing daily water-level fluctuations in the Peninsula well with local barometric and sea-level fluctuations, February 12 - 28, 1970.
Water levels in wells that tap coastal artesian aquifers often fluctuate naturally in response to atmospheric, earth, and ocean tides (Bredehoeft, 1967; Robinson and Bell, 1971). This response is usually due to the cyclic loading of the aquifer, which is theoretically transmitted undiminished through the overlying confining layer. Therefore, as the load on the confined water and aquifer skeleton is increased, the water level in the well rises because of increased pressure in the aquifer. The amplitude of the fluctuation in the well is dampened because the aquifer, having some rigidity, supports part of the changing load, and only the remainder of this change is reflected in the pressure of the water.

As previously stated, the water level in the Peninsula well is slightly higher than the pressure head in the Boulder Zone, owing to operationally-caused differences in the salinity and temperature, hence, the density of the water column. However, short-term natural cyclic fluctuations of the water level, such as tidal effects, are not significantly affected by this difference in density because the additions to and subtractions from the height of the water column are made up of water from the Boulder Zone.

**ATMOSPHERIC TIDE**

Since the cyclic variations at atmospheric pressure are related to the sun’s transit they have a resonance period of 12 hours. The atmospheric pressure is greatest at about 10:00 a.m. and 10:00 p.m. and least at about 4:00 a.m. and 4:00 p.m. The variations are primarily the result of the combined effects of the sun’s gravitational attraction and solar heating, with solar heating being the major component. As previously shown in the section on long-term atmospheric effects, the water level in an artesian well rises during declining atmospheric pressure and declines during rising atmospheric pressure. Barometric fluctuations in wells, therefore, have a semidiurnal component in which the minima occur at about 10:00 a.m. and 10:00 p.m. and the maxima at about 4:00 a.m. and 4:00 p.m.

The effect of atmospheric tide on the water level in the Boulder Zone appears to be minimal, according to figures 7 and 8. The water level seems to be fluctuate chiefly in response to ocean tides. On the basis of a barometric efficiency of 70 percent and the range of semidiurnal variations in barometric pressure, the corresponding water-level fluctuations in the well is less than 0.05 foot.
The gravitational attraction of the moon and sun causes the ocean tides and earth tides, or cyclic bulging up of the ocean and land. The periods of the ocean and earth tides are both about 12.4 hours, but the effect of earth tide on the level in wells that penetrate artesian aquifers is opposite to that of ocean tide in that the water levels in these wells are low during the moon's transit. Details of the effects of earth tides on well-aquifer systems are described in a report by J. D. Bredehoeft (1967).
Earth tide, although relatively small, reinforces or diminishes the apparent effect of oceanic tides on the water level in the well, depending upon the phase relationship. Earth tides are alternately in and out of phase with the semidiurnal atmospheric fluctuations in the well, augmenting them near the new and full moon phases and counteracting them near first and third quarter phases.
Daily water-level fluctuations in the Peninsula well are chiefly caused by the semidiurnal lunar tide (fig. 7). The period of the lunar tides is 12.4 hours; therefore, high and low tides arrive about 0.8 hour later each day. A comparison of the water-level fluctuations in the well with the tidal fluctuations in Biscayne Bay at Coconut Grove and in the ocean at Miami Beach shows that water-level peaks and troughs seem to be related to the high and low tides. Diurnal inequalities are indicated by unequal tides each day. Water-level fluctuations in the well ranged from 0.2 to 0.4 foot, despite the fact that the well is about 6 miles inland from Biscayne Bay, the nearest body of tidal water.

The amplitude of the water-level fluctuations is only 10 to 20 percent that of the oceanic tides, depending on the tides at the tidal station used for the comparison. The amplitude of the fluctuations in the well seems to be about 17 percent of that in Biscayne Bay and about 13 percent of that in the ocean at Miami Beach. These values, however, do not represent the true ratios, hence tidal efficiency (TE), of the aquifer. According to Jacob (1950, p. 331 - 332), the sum of the barometric and tidal efficiencies equals unity, that is

\[ BE + TE = 1 \]  

The barometric efficiency was determined to be 0.70 (p. 28), therefore the tidal efficiency would be 0.30, according to this relationship.

The comparison of the semidiurnal water-level fluctuations in the Peninsula well with semidiurnal tides in Biscayne Bay at Coconut Grove (fig. 7) indicated that water-level peaks and troughs led corresponding tides by about ½ hour and 1¾ hours, respectively, or by an average of about 3/4 hour. Schneider (1969) showed that tides in Biscayne Bay usually lagged ocean tides at Miami Beach by more than 1 hour. On February 7, tides at Miami Beach (Miami Harbor Entrance) led corresponding water-level fluctuations in the Peninsula well from ¾ to 1 hour, or by an average of 40 minutes (from preliminary tide data, U. S. Dept. of Commerce, August 1971).

Thus the well fluctuations led tide fluctuations everywhere in Biscayne Bay, but lagged tides at Miami Beach. This relation is important with respect to the area of aquifer loading and the degree of hydraulic interconnection of the Boulder Zone with the Straits of Florida.

The relation between tides in the well and tides in Biscayne Bay suggested that well tides were not related to loading in Biscayne Bay. However, an analysis of the fluctuations by H. H. Cooper (written commun., March 15, 1973) indicated that the anomaly was caused chiefly by the added effect of earth tide. Therefore, the semidiurnal water-level fluctuation in the well was chiefly the
result of at least two out-of-phase components, namely earth tide and ocean tide.

Semidiurnal atmospheric fluctuations, although small, are also out of phase with the ocean tide and earth tide components and would therefore affect the relation between tides in the well and tides in Biscayne Bay. However, by selecting periods during which barometric changes were minimal this factor was eliminated.

According to Cooper, the troughs of the earth tide component should nearly coincide with the overhead transit of the moon, and the troughs of the ocean tide component should lag the earth tide component both because ocean tide is out of phase with earth tide and the velocity of the incoming ocean tide component is reduced by the impedance of the aquifer. The water-level fluctuation in the well would therefore be out of phase with each component, resulting in water-level troughs that lag the troughs due to the earth tide component and that precede the troughs due to the ocean tide component.

For convenience in the water-level analysis the semidiurnal fluctuation in the well was assumed to be due chiefly to the sum of two out-of-phase components, ocean tide and earth tide, and that the two components were purely sinusoidal with a period of 12.4 hours (360°).

The times of the moon's transit, according to the Nautical Almanac 1970-71 (U. S. Naval Observatory, 1968 and 1969), were compared with water-level troughs in the well during November 25 - 27, 1970 and February 5 - 7, 1971, and the troughs seemed to lag the transits by 2½ hours (72.5°). Water level troughs in the well were assumed to precede the arrival of low tide in Biscayne Bay at Coconut Grove by about 3/4 hours (21.7°) on the average. The time required for the tidal effect to travel from the coastline to the well was determined indirectly by an iterative process involving estimates of the hydraulic diffusivity of the aquifer, as will be discussed later in the section dealing with the aquifer characteristics. The value thus obtained was 1/3 hour (9.8°).

The amplitudes of the ocean tide and earth tide components also were determined indirectly by the iterative process involving estimates of the hydraulic diffusivity of the aquifer and by trigonometric equations suggested by Cooper.

The analysis of water-level fluctuations is expressed as:

\[
\text{Observed water-level fluctuation in the well = fluctuation due to earth tide + fluctuation due to ocean tide.}
\]
The resultant equation for the harmonic analysis is

\[ 0.12 \sin (\Omega t) = 0.065 \sin (\Omega t + 72.5^\circ) + 0.118 \sin (\Omega t + 31.5^\circ) \]  

(3)

where \( \Omega = \) the angular frequency \( = \frac{360^\circ}{\tau} = \frac{360^\circ}{12.4} \) degrees/hour

\( t = \) elapsed time with respect to an initial reference in hours.

\( \tau = \) period of fluctuation - 12.4 hours.

The range of fluctuation in the well was 0.240 foot; the range of the earth tide component was 0.130 foot; and the range of the ocean tide component was 0.236 foot. The components of equation 3 are plotted on figure 9.

**WATER LEVEL FLUCTUATIONS IN THE UPPER PART OF THE FLORIDAN AQUIFER**

The pressure head in the principal artesian water-bearing zone of the Floridan aquifer was measured at about 41 feet above msl by a mercury manometer during February 10 - 11, 1971. The pressure head representing the composite pressures of artesian zones exposed to the annular space between casings (between 545 and 1,678 feet) seemed to fluctuate slightly in response to semidiurnal changes in atmospheric pressure (fig. 8). The difference between the water level in the principal artesian zone and the water level in the Boulder Zone is attributed by the author chiefly to the difference in fluid density.

The apparent lack of response of the pressure head in the Floridan to ocean tide is due to the great disparity between hydraulic characteristics of the principal artesian zone and those of the Boulder Zone. The difference in response of the two aquifer systems coupled with geologic considerations, suggests the following: (1) the Boulder Zone and the principal artesian zone of the Floridan aquifer may function as hydraulically separate systems, (2) a hydraulic interconnection may exist between both the Boulder Zone and the Straits of Florida, (3) a hydraulic interconnection may exist between the principal artesian zone of the Floridan aquifer and the Straits of Florida, (4) permeabilities in the horizontal direction exceed permeabilities in the vertical direction by an appreciable amount, and (5) the transmissivity of the Boulder Zone is much greater than the transmissivity of the principal artesian zone of the Floridan aquifer.
HYDRAULIC CHARACTERISTICS OF THE BOULDER ZONE

RELATION BETWEEN AQUIFER CHARACTERISTICS AND SHORT-TERM CYCLIC WATER-LEVEL FLUCTUATIONS

The hydraulic characteristics of an aquifer can be determined from short-term natural water-level fluctuations in a well. Ferris (1951) described a method that related cyclic (sinusoidal) water-level fluctuations in wells and nearby surface-water bodies to hydraulic diffusivity (the ratio of transmissivity to storage coefficient $T/S$, or, equivalently, the ratio of hydraulic conductivity to specific storage $K/S_s$ of an aquifer). The method assumes that the aquifer is homogeneous, of uniform thickness, and of great areal extent. Furthermore, the aquifer is assumed to be bounded along a straight line on one side by a surface-water body. Within the aquifer, water storage is assumed to change instantaneously with and at a rate proportional to the change in head. Tidal fluctuations are, therefore, transmitted horizontally through the aquifer from the surface water-aquifer contact. The fluctuation decreases exponentially with time and distance from the source.

Carr and Van Der Kamp (1969) modified the method to permit the determination of hydraulic conductivity and specific storage for a confined aquifer underlying the ocean. Tidal fluctuations in inland wells are induced by the inland migration through the aquifer of the pressure waves produced by the cyclic loading of the ocean. Fluctuations of water levels in the inland wells are less than those of the pressure head in the part of the aquifer beneath the sea, owing to the energy losses that accompany the to-and-fro movement of water in the aquifer. Later Van Der Kamp (1972) showed that the tidal efficiency at the coastline would be about one-half the loading efficiency; and that the loading efficiency is essentially the tidal efficiency expressed in equation 2.

The dampening of a oceanic tidal fluctuation in an artesian aquifer during transit from the shoreline to an inland well (Ferris, 1962, p. 133, equation 63) can be written

$$h_w = h_o e^{-x \sqrt{\pi S_s / T}}$$

(4)

when $h_w$ = Fluctuation in the well at distance $x$ from the shoreline

$h_o$ = fluctuation in the aquifer at the shoreline ($x = 0$)
The time required for a given oceanic tidal fluctuation to travel through an artesian aquifer from the shoreline to an inland well (Ferris, 1962, p. 134, equation 67) can be written

\[ t_1 = \frac{x \sqrt{T \tau S}}{2 \pi T} \]  

where \( t_1 \) = the time lag at distance x from the shoreline.

**HYDRAULIC DIFFUSIVITY**

Hydraulic diffusivity is defined as the conductivity of the aquifer when the unit volume of water moving is that involved in changing the head a unit amount in a unit volume of aquifer (Lohman, 1972, p. 8). As stated earlier it is essentially the ratio of transmissivity (T) to the storage coefficient (S) or the ratio of hydraulic conductivity (K) to specific storage (Ss).

\[ \text{Hydraulic diffusivity} = \frac{T}{S} = \frac{K}{S_s} \]

Equations 4 and 5, the stage-ratio and time-lag equations of Ferris, were rearranged and solved for diffusivity (T/S):

\[ \frac{T}{S} = \frac{K}{S_s} = \frac{x^2 \pi}{\tau} \frac{1}{\frac{\ln h_o}{h_w}} \]  

\[ \frac{T}{S} = \frac{K}{S_s} = \frac{x^2 \tau}{4 \pi t_1^2} \]

The hydraulic diffusivity of the Boulder Zone was computed by a method using equations 7 and 8, trigonometric equations suggested by Cooper to separate ocean-tide and earth-tide components of the water-level fluctuations in
the Peninsula well, and the relation between tidal effects at the shoreline and tidal efficiency that was proposed by Van Der Kamp (1972).

On the basis of Van Der Kamp's theory of tidal efficiency at the shoreline, the tidal fluctuation \( h_0 \) in the Boulder Zone at the shoreline was calculated from tidal fluctuations in Biscayne Bay at Coconut Grove during November 25 - 27, 1970 and February 5 - 8, 1971. The tidal fluctuation in Biscayne Bay at Coconut Grove was 1.86 feet on the average; therefore, the fluctuation in a well tapping the Boulder Zone at the shoreline \( h_0 \) would be 0.28 foot on the premise that the tidal efficiency at the shoreline is about half the loading efficiency (0.30).

The distance \( x \) from the shoreline to the Peninsula well is 6 miles. The analysis of water-level fluctuations in the well and tide fluctuations at Coconut Grove indicated that the ocean tide component lagged the well component by 3/4 hour (21.7°) in addition to the time required for the tidal fluctuation to travel from the shoreline through the aquifer to the well, which is equivalent to \( t_1 \) in equation 5.

By assuming values of \( t_1 \) it was possible to calculate phase and amplitude relationships and approximate diffusivity values using the time-lag and stage-ratio equations. The substitution or iterative process continued until a diffusivity value was obtained that satisfied both equations and the phase-amplitude relationship in figure 9. The results of the iterative process yielded a diffusivity of \( 2.1 \times 10^{11} \) ft\(^2\)/day.

**SPECIFIC STORAGE AND HYDRAULIC CONDUCTIVITY**

The specific storage, \( S_s \) is the volume of water released from or taken into storage per unit volume of aquifer per unit change in head. Hydraulic conductivity, \( K \), is the volume of water that will move in a unit time under a unit hydraulic gradient through a unit area of aquifer measured at right angles to the direction of flow.

The specific storage was calculated for porosities ranging from 10 to 90 percent by the following equation (Bredehoeft, 1967, p. 3083; Carr and Van Der Kamp, 1969, p. 1023):

\[
S_s = \frac{\theta \beta \gamma}{BE}
\]  

(9)
Figure 9  Graph showing earth-tide and ocean-tide components of well fluctuation.
where $\theta$ = porosity

$\beta$ = compressibility of water ($2.23 \times 10^{-8} \text{ ft}^2/\text{lb at } 60^\circ\text{F}$)

$\gamma$ = specific weight of water (64 lbs/ft$^3$ for salt water)

$BE$ = Barometric efficiency (0.70)

Therefore

$$S_s = \frac{\theta (2.23 \times 10^{-8}) (64) \cdot \text{ft}^2 \cdot \text{lb}}{7.0 \times 10^{-1} \cdot \text{lb} / \text{ft}^3}$$

$$= \theta (2.04 \times 10^{-6}) \text{ ft}^{-1};$$

Based on the relationships in equation 6, the hydraulic conductivity was calculated for porosities ranging from 10 to 90 percent by the following equation:

$$K = \text{hydraulic diffusivity} \times S_s,$$

(10)

substituting

$$K = (2.1 \times 10^{11}) (2.04 \times 10^{-6}) \frac{\theta \cdot \text{ft}^2 \cdot 1}{\text{day} \cdot \text{ft}}$$

$$= 4.28 \times 10^5 \theta \text{ ft/day}$$

The results of the computations are shown in table 1. The values of $K$ and $S_s$ vary widely, depending upon the porosity. Generally, the connected porosity of very permeable limestone aquifers, like the Biscayne aquifer, which is about 100 feet thick, ranges from 20 to 30 percent. The caliper-flowmeter log of the Peninsula well showed that the main water-bearing zone occurs between 2,930 and 2,945 feet. The porosity of the 15-foot zone is probably higher than that of the Biscayne aquifer.

In the author’s opinion a reasonable estimate of porosity for the 15-foot thick zone is 50 percent. The specific storage and hydraulic conductivity values corresponding to a 50 percent porosity (table 1) are $1.0 \times 10^{-6}$ ft$^{-1}$ and $2.1 \times 10^5$ ft per day, respectively. If, on the other hand, the thickness of the water-bearing zone is greater than 15 feet one would expect a lesser value for porosity.
TABLE 1
Estimated values of hydraulic conductivity and specific storage
for porosities ranging from 10 to 50 percent.

$\theta$ , percent porosity

$S_s$, specific storage, ft$^{-1}$

$K$, hydraulic conductivity, ft day$^{-1}$

<table>
<thead>
<tr>
<th>$\theta$</th>
<th>$K$</th>
<th>$S_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>$4.3 \times 10^4$</td>
<td>$2.0 \times 10^{-7}$</td>
</tr>
<tr>
<td>20</td>
<td>$8.6 \times 10^4$</td>
<td>$4.1 \times 10^{-7}$</td>
</tr>
<tr>
<td>30</td>
<td>$1.3 \times 10^5$</td>
<td>$6.1 \times 10^{-7}$</td>
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<tr>
<td>40</td>
<td>$1.7 \times 10^5$</td>
<td>$8.2 \times 10^{-7}$</td>
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<tr>
<td>50</td>
<td>$2.1 \times 10^5$</td>
<td>$1.0 \times 10^{-6}$</td>
</tr>
<tr>
<td>60</td>
<td>$2.6 \times 10^5$</td>
<td>$1.2 \times 10^{-6}$</td>
</tr>
<tr>
<td>70</td>
<td>$3.0 \times 10^5$</td>
<td>$1.4 \times 10^{-6}$</td>
</tr>
<tr>
<td>80</td>
<td>$3.4 \times 10^5$</td>
<td>$1.6 \times 10^{-6}$</td>
</tr>
<tr>
<td>90</td>
<td>$3.9 \times 10^5$</td>
<td>$1.8 \times 10^{-6}$</td>
</tr>
</tbody>
</table>

A method suggested by Bredehoeft (1967, p. 3083), was also used to estimate specific storage and porosity. The method assumes that Poisson's ratio for the aquifer is known. The specific storage was calculated to be $1.4 \times 10^{-7}$ ft$^{-1}$, based on the assumption that the earth tide component (0.13 ft) at the well is equal to the change in head produced by the tidal dilatation at the earth's surface ($\Delta_t$) and on other assumptions used by Bredehoeft (1967, eq. 25). The porosity of the aquifer was calculated to be 7 percent using equation 9; and the hydraulic conductivity was calculated to be $3.0 \times 10^4$ ft/day$^{-1}$ using equation 10.

A comparison of values of $K$ and $S_s$ for 50 percent porosity (table 1) with the values derived by the Bredehoeft method suggests that a 15-foot thick aquifer with 50 percent porosity is hydraulically equivalent to a 105-foot thick aquifer with 7 percent porosity. Therefore it is conceivable that the true values of hydraulic conductivity and specific storage for the aquifer could range between those for 7 percent porosity and 50 percent porosity, depending upon the thickness of the permeable zone. In the author's opinion, the physical situation supports the values for a 15-foot thick aquifer.
TRANSMISSIVITY AND STORAGE COEFFICIENT

The transmissivity (T) and the storage coefficient (S) of the aquifer, or zone, are related respectively to the hydraulic conductivity (K) and the specific storage (S_s) by the equations

\[ T = mK \]  \hspace{1cm} (11)
\[ S = mS_s \]  \hspace{1cm} (12)

where m is thickness. These equations assume that K and S_s are uniform. Computed values of T and S for porosities ranging from 10 to 90 percent are presented in Table 2.

<table>
<thead>
<tr>
<th>( \theta )</th>
<th>T ( \text{ft}^2 \text{day}^{-1} )</th>
<th>S ( \text{dimensionless} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>6.4 ( \times 10^5 )</td>
<td>3.1 ( \times 10^6 )</td>
</tr>
<tr>
<td>20</td>
<td>1.3 ( \times 10^6 )</td>
<td>6.1 ( \times 10^6 )</td>
</tr>
<tr>
<td>30</td>
<td>1.9 ( \times 10^6 )</td>
<td>9.2 ( \times 10^6 )</td>
</tr>
<tr>
<td>40</td>
<td>2.6 ( \times 10^6 )</td>
<td>1.2 ( \times 10^5 )</td>
</tr>
<tr>
<td>50</td>
<td>3.2 ( \times 10^6 )</td>
<td>1.5 ( \times 10^5 )</td>
</tr>
<tr>
<td>60</td>
<td>3.9 ( \times 10^6 )</td>
<td>1.8 ( \times 10^5 )</td>
</tr>
<tr>
<td>70</td>
<td>4.5 ( \times 10^6 )</td>
<td>2.1 ( \times 10^5 )</td>
</tr>
<tr>
<td>80</td>
<td>5.1 ( \times 10^6 )</td>
<td>2.4 ( \times 10^5 )</td>
</tr>
<tr>
<td>90</td>
<td>5.8 ( \times 10^6 )</td>
<td>2.8 ( \times 10^5 )</td>
</tr>
</tbody>
</table>

1 Based on a 15-foot thickness of aquifer.

The caliper-flowmeter log of the Peninsula well showed that the cavernous water-bearing zone extends in depth from 2,930 to about 2,945 feet, a thickness of 15 feet. If the porosity is assumed to be about 50 percent, then the transmissivity and storage coefficient would be 3.2 \( \times 10^6 \) \( \text{ft}^2 \text{per day} \) and 1.5 \( \times 10^5 \), respectively. The same values of T and S would apply for a 105-foot
thick water-bearing zone with a 7 percent porosity, a hydraulic conductivity of $3.0 \times 10^4 \text{ ft/day}$, and a specific storage of $1.4 \times 10^{-7} \text{ per foot}$.

Lohman (1972, p. 52) suggests that storage coefficient can be estimated by multiplying the aquifer thickness in feet times $10^{-6} \text{ n}^{-1}$. The value for a 15-foot thick aquifer would be $1.5 \times 10^{-5}$ which is equivalent to that computed on the basis of the hydraulic diffusivity ($2.1 \times 10^{11} \text{ ft}^2 \text{ day}^{-1}$) and a 15-foot thick water-bearing zone with a 50 percent porosity. Based on a specific capacity test during the early stage of well development, J. I. Garcia-Bengochea (oral commun., November, 1971) estimated that the transmissivity was about $1.6 \times 10^5 \text{ ft}^2 \text{ day}^{-1}$, which compares reasonably well with the author’s estimate of $3.2 \times 10^5 \text{ ft}^2 \text{ day}^{-1}$.

According to data from oil test wells in southeast Florida, cavernous zones might extend in depth from about 2,900 feet to perhaps 3,500 feet. The hydraulic characteristics (T and S) for the 15-foot water-bearing zone are, in the author’s opinion, representative of the geologic conditions at the Peninsular well, but the same values could apply to a thicker zone with lower porosity. The estimates herein will be improved upon as more reliable data are obtained from pumping tests.

**SUMMARY AND CONCLUSIONS**

The natural water-level fluctuations in the 2,947-foot-deep Peninsula Utilities disposal well open to the Boulder Zone, near Miami, Florida, are caused largely by atmospheric fluctuations, earth tides, and ocean tides. The water level ranged from 7.1 to 3.4 feet above mean sea level as a result of operationally-caused variations in the density of the water column. The pressure head of the Boulder Zone at the 2,947-foot depth in terms of equivalent seawater density fluctuates at or near present sea level.

Semidiurnal natural fluctuations in the well are caused chiefly by loading due to earth tides and tides in Biscayne Bay. The tidal efficiency of the Boulder Zone is about 30 percent, and the barometric efficiency is about 70 percent. An analysis of cyclic water-level fluctuations in the well suggests that the earth-tide and ocean-tide components are about 104 degrees out of phase. The range of fluctuation in the well was 0.24 foot; the component due to ocean tide was 0.236 foot; and the component due to earth tide was 0.130 foot. The amplitude and phase relationships are apparently responsible for tides in Biscayne Bay lagging corresponding tides in the well tides by 3/4 hour. The geology, water quality, and temperature gradient suggest a hydraulic connection between the Boulder Zone and the Straits of Florida about 35 miles to the east of Miami.
The effect of tidal loading on the principal artesian water-bearing zone of the Floridan aquifer was not apparent in measurements of pressure head at the well, although this aquifer also underlies Biscayne Bay and crops out in the Florida Straits. The difference in response of the two aquifer systems to tides at the Peninsula well suggests that they function as hydraulically separate systems (although evidence is inconclusive), that horizontal permeabilities significantly exceed vertical permeabilities, and that the transmissivity of the Boulder Zone is much greater than the transmissivity of the principal artesian water-bearing zone of the Floridan aquifer.

The hydraulic diffusivity of the Boulder Zone was computed to be $2.1 \times 10^{11}$ ft$^2$/day. The hydraulic conductivity and specific storage of the 15-foot permeable zone are estimated to be $2.1 \times 10^5$ ft/day and $1.0 \times 10^{-7}$ per ft, respectively, based on an assumed porosity of 50 percent. The transmissivity and storage coefficient of the Boulder Zone are estimated to be $3.2 \times 10^6$ ft$^2$/day and $1.5 \times 10^{-5}$, respectively. Pumping tests and precise water-level and density data are needed to determine deep circulation patterns accurately and the ultimate direction and rate of movement of injected fluids.
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