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Tide-induced groundwater level fluctuation in coastal aquifers bounded by L-shaped coastlines

Hailong Li,[1,2] Jiu Jimmy Jiao,[1] Mario Luk,[3] and Kayiu Cheung[1]

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[1] This paper presents an analytical solution to describe tidal groundwater level fluctuations in an aquifer bounded by two water-land boundaries that form a right angle. The aquifer configuration can represent the situation of a right-angle corner in an island when the tidal boundaries are the same along both boundaries, or the estuary situation if the amplitude damps with distance along one boundary (river) and the amplitude along the other boundary (sea) does not change spatially. The two-dimensional flow equation subject to periodic boundary conditions is changed into a time-independent elliptic problem using complex transform. The elliptic problem is then solved using the Green's function method. *Li et al.* [2000] obtained a nonperiodic solution to the same problem with an additional hypothetical initial condition. Their solution includes a time-dependent integral. The solution presented here is a significant improvement over theirs in the sense that the solution is periodic and computationally efficient since the integral involved is independent of time. An approximate solution without integral is also derived. An error analysis and a case study in Hong Kong demonstrate that the approximate solution is adequate for most field problems. *INDEX TERMS:* 1829 Hydrology: Groundwater hydrology; 1832 Hydrology: Groundwater transport; 1899 Hydrology: General or miscellaneous; 3384 Meteorology and Atmospheric Dynamics: Waves and tides; *KEYWORDS:* periodic groundwater flow, analytical solution, coastal aquifer, sea tide, L-shaped coastline

1. Introduction

[2] Tide-induced groundwater level fluctuations are an interesting natural phenomenon in coastal aquifers. Since the 1950s, research in this field has attracted much attention from hydrogeologists. *Jacob* [1950] derived an equation for the tidal groundwater fluctuation in a single coastal confined aquifer. This equation has been widely used [e.g., *Carr and van der Kamp*, 1969; *Serfes*, 1991]. Meanwhile, other analytical solutions related to tidal fluctuations were derived for coastal aquifers. For example, *Nielsen* [1990] discussed the effects of a slope boundary on tidal water table fluctuations in unconfined coastal aquifers; *Sun* [1997] developed an analytical solution in an estuary using a two-dimensional tidal loading boundary condition. *Jiao and Tang* [1999], *Li and Jiao* [2001a, 2001b], and *Tang and Jiao* [2001] derived analytical solutions for multilayer coastal leaky aquifer systems. All these studies, however, assumed that the coastline is straight.

[3] In reality, coastlines are very irregular and full of inlets, bays, and headlands and cannot always be, even approximately, regarded as straight lines. Coastlines that bend at a right angle (hereinafter referred to as "L-shaped coastlines"; see Figure 1) are an idealized situation, but this situation is believed to be of some practical use. For example, in most estuaries, which are frequently studied coastal areas, the coastlines are often approximately L-shaped [e.g., *Cheng and Chen*, 2001; *Harrison and Boon*, 1972; *Gregg*, 1966; *Li et al.*, 2000]. Other aquifers with L-shaped tidal boundaries are given by *Carr and van der Kamp* [1969], *Maas and De Lange* [1987], and *Liu* [1996].

[1] Department of Earth Sciences, University of Hong Kong, Hong Kong, People's Republic of China.
[2] Also at Institute of Biomathematics, Anshan Normal College, Anshan, Liaoning, People's Republic of China.
[3] Provisional Airport Authority, Hong Kong, People's Republic of China.

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[4] As the first work to address the impact of the coastline shape on tidal groundwater fluctuation in coastal aquifers, *Li et al.* [2000] derived a spatially two-dimensional analytical solution in an unconfined aquifer with L-shaped coastlines. They obtain their solution by solving an initial-boundary value problem using the Green's function method. Their solution is in terms of complex integration of time. Because of the hypothetical initial condition, their solution is not periodic. In order to obtain an approximate periodic solution, the influence of the initial condition on the solution has to be eliminated by choosing a sufficiently large integral interval of time. The integral in their solution has to be evaluated numerically. Because of these the computational effort of their solution is great.

[5] In this paper, attempts are made to derive more simple analytical solutions to the model by *Li et al.* [2000]. Instead of the initial-boundary value problem that they considered, only a boundary value problem is solved to avoid the influence of the hypothetical initial condition on the solution, and a periodic solution is directly obtained. This solution is then compared with the solution of *Li et al.* [2000]. Discussion is made on the advantages of the solution over that of *Li et al.* [2000]. A more simple approximation of the solution, which can be evaluated even with a calculator, is also given. The accuracy of the approximate solution is analyzed. Finally, both the new solution and Jacob's solution are used to estimate the parameter of an L-shaped, deep unconfined aquifer formed by the reclamation fill around Lam Chau Island, which is now a part of the Chek Lap Kok Airport, Hong Kong SAR (Special Administration Region), People's Republic China. The aquifer parameters estimated based on the two solutions are compared and discussed.

2. Mathematical Model and Analytical Solution

[6] Assume that an aquifer lies in the first quadrant with the positive parts of both the x and y axes being the L-shaped coastlines (Figure 1). The aquifer is homogeneous, horizontal, with

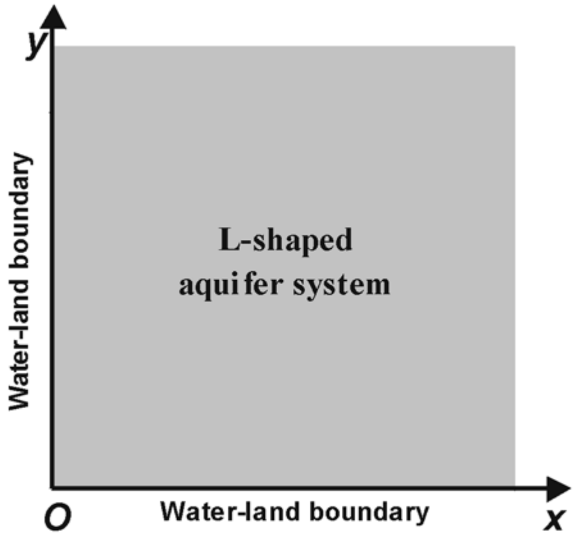

Figure 1. A coastal aquifer with L-shaped coastlines.

constant thickness, and connected directly to the tidal water along the coastlines. The water-land boundary is vertical. In case that the aquifer is unconfined, it is assumed that the ratio of the tidal amplitude to the average aquifer thickness is small enough so that Boussinesq's equation can be linearized [see, e.g., *Bear*, 1972]. Then the groundwater flow equation in the L-shaped aquifer is [*Bear*, 1972; *Li et al.*, 2000]

$$\frac{S}{T}\frac{\partial h}{\partial t}=\frac{\partial^2 h}{\partial x^2}+\frac{\partial^2 h}{\partial y^2}, \qquad -\infty < t < +\infty, \quad x, y > 0, \tag{1}$$

where S is the storativity (dimensionless) of the aquifer; T is the transmissivity [L^2 T^{-1}] of the aquifer; and $h(x,y,t)$ is the groundwater head [L] for a confined aquifer or the water table for an unconfined aquifer. On the side ($y > 0$, $x = 0$) which represents the boundary in the estuary, the tidal attenuation is considered by [*Li et al.*, 2000]

$$\begin{aligned} h(x,y,t)\big|_{x=0,\,y>0} &= A\exp(-\kappa_{er}y)\cos(\omega t-\kappa_{ei}y+c)\\ &= \mathrm{Re}\left(e^{-\kappa_e y+i(\omega t+c)}\right), \end{aligned} \tag{2}$$

where y denotes the distance along the estuary from the entry; A and ω are the tidal amplitude and frequency, respectively; $\kappa_{er} \geq 0$ and $\kappa_{ei} \geq 0$ are the amplitude damping coefficient and wave number of the tidal wave in the estuary, respectively; $\kappa_e = \kappa_{er} + i\kappa_{ei}$; and c is the phase shift. The datum of the hydraulic head of the aquifer is set to be the mean sea level. On the other side ($y = 0$, $x > 0$), which represents the ocean-land boundary, the spatially constant tidal boundary condition

$$h(x,y,t)_{y=0,\,x>0}= A\cos(\omega t+c) \tag{3}$$

is used. In inland places far from the origin, no-flow boundary condition

$$\lim_{x\to\infty}\frac{\partial h}{\partial x}(x,y,t)=\lim_{y\to\infty}\frac{\partial h}{\partial y}(x,y,t)=0 \tag{4}$$

is used. The derivation of the solution $h(x,y,t)$ to the boundary value problem (1), (2), (3), and (4) is presented in Appendix A. The result reads

$$\begin{aligned} h(x,y,t) = A\mathrm{Re}[(&I(ax,ay;1+i,1+i)+I(ay,ax;m+in,1+i)\\ &+e^{-(1+i)ay}+e^{-\kappa_e y-(m+in)ax})e^{i\omega t+c}], \end{aligned} \tag{5}$$

where

$$\begin{aligned} I(\xi,\eta;\mu,\lambda) = -\frac{\lambda^2\xi}{\pi}\int_0^\infty & e^{-\mu\tau}\\ &\cdot\left(\frac{K_1(\rho(\xi,\eta-\tau;\lambda))}{\rho(\xi,\eta-\tau;\lambda)}-\frac{K_1(\rho(\xi,\eta+\tau;\lambda))}{\rho(\xi,\eta+\tau;\lambda)}\right)d\tau, \end{aligned} \tag{6}$$

with $K_n(\rho)$ being the modified second-kind Bessel function of nth order and

$$\rho(\xi,\eta;\lambda)=\lambda\sqrt{\xi^2+\eta^2}; \tag{7}$$

the two dimensionless, positive constants m and n are defined as

$$m^2=\sqrt{\left(1-\frac{\kappa_{er}\kappa_{ei}}{a^2}\right)^2+\left(\frac{\kappa_{er}^2-\kappa_{ei}^2}{2a^2}\right)^2}-\left(\frac{\kappa_{er}^2-\kappa_{ei}^2}{2a^2}\right)^2, \tag{8}$$

$$n^2=\sqrt{\left(1-\frac{\kappa_{er}\kappa_{ei}}{a^2}\right)^2+\left(\frac{\kappa_{er}^2-\kappa_{ei}^2}{2a^2}\right)^2}+\left(\frac{\kappa_{er}^2-\kappa_{ei}^2}{2a^2}\right)^2, \tag{9}$$

and a is the aquifer's tidal propagation parameter [L^{-1}],

$$a=\sqrt{\frac{\omega S}{2T}}. \tag{10}$$

3. Discussion of the Solution

3.1. Comparison With the Solution of *Li et al.* [2000]

[7] *Li et al.* [2000] obtained a nonperiodic solution to (1) subject to the boundary conditions (2)–(4) and an additional

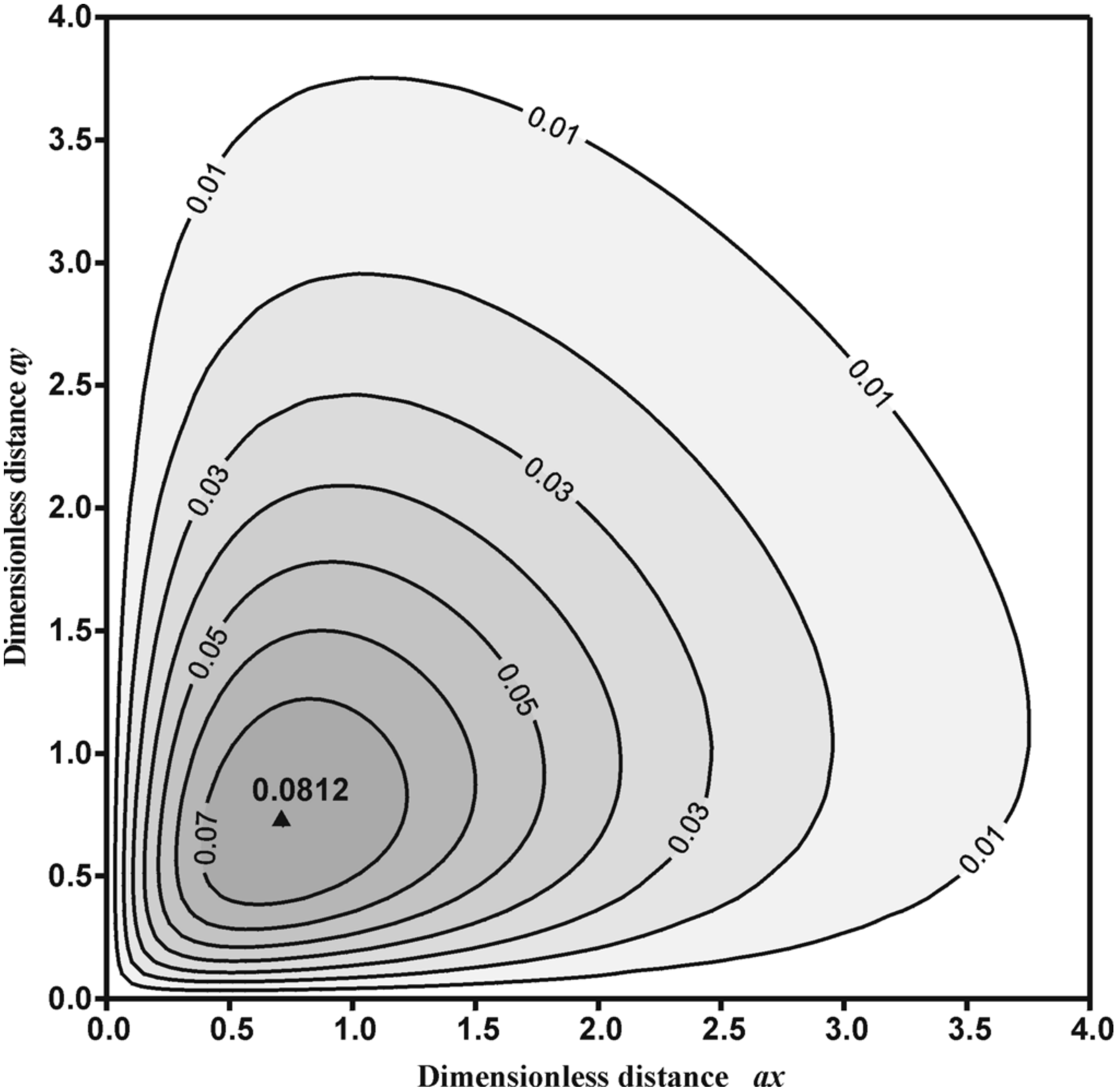

Figure 2. Maximum error distribution $R(ax,ay)$ of the simple approximate solution (11) when both κ_{er} and κ_{ei} vanish.

hypothetical initial condition. Their solution includes complex integration of time. In order to eliminate the influences of the hypothetical initial condition on the solution, the integral interval of time must be sufficiently large. Their solution tends to be periodic when t approaches infinite. Compared with their solution, the solution (5) has advantages such as simplicity and periodicity because it includes no initial condition. Although both solutions include integrals that have to be evaluated numerically, the integral in the solution (5) is independent of time, and hence it can be calculated more efficiently. Once the integral in (5) is evaluated, the solution (5) can determine the tidal groundwater level for any time; the solution of *Li et al.* [2000], however, has to reevaluate the integral for different time because the integral depends on time.

3.2. Approximate Simplification of Solution (5)

[8] The solution (5) can be further simplified approximately into

$$h_{\text{approx}}(x,y,t) = A\text{Re}\left[\left(-e^{-(1+i)ay-(m+in)ax} + e^{-(1+i)ay} + e^{-\kappa_e y-(m+in)ax}\right)e^{i\omega t+c}\right]. \tag{11}$$

Equation (11) is an approximate solution of (1), (2), (3), and (4) because it only satisfies the boundary condition (2), (3), and (4) but does not satisfy the differential equation (1). Let

$$R_r = \text{Re}\left[I(ax,ay;1+i,1+i) + I(ay,ax;m+in,1+i) + e^{-(1+i)ay-(m+in)ax}\right], \tag{12a}$$

$$R_i = \text{Im}\left[I(ax,ay;1+i,1+i) + I(ay,ax;m+in,1+i) + e^{-(1+i)ay-(m+in)ax}\right]. \tag{12b}$$

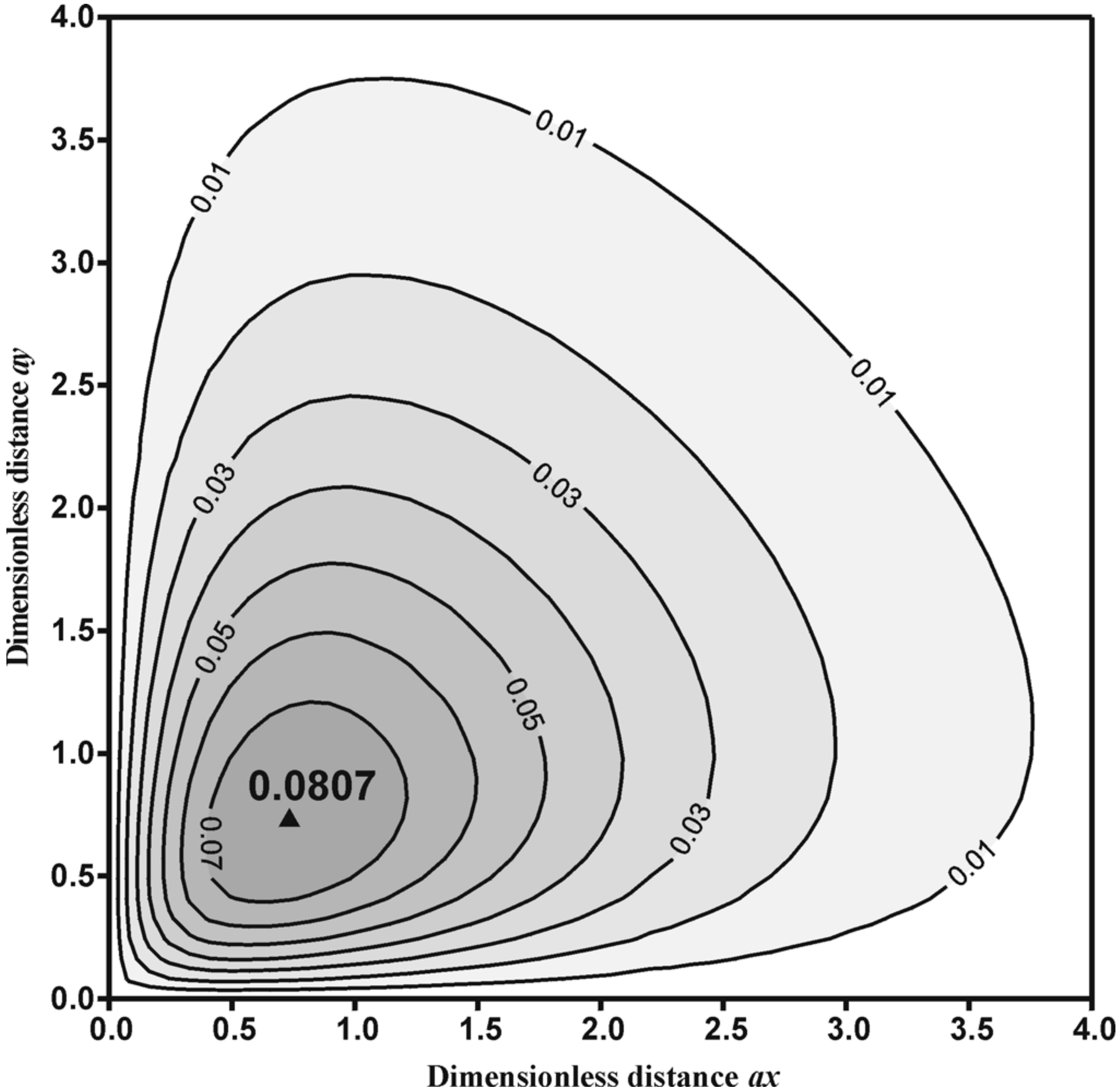

Figure 3. Maximum error distribution $R(ax,ay)$ of the simple approximate solution (11) when $\kappa_{er} = \kappa_{ei} = 0.1a$.

Then the spatial maximum error distribution of the approximate solution (11) relative to the tidal amplitude A is given by

$$R(ax,ay) \stackrel{\text{def}}{=} \max_t \quad |h(x,y,t) - h_{\text{approx}}(x,y,t)|/A = \sqrt{R_r^2 + R_i^2}. \tag{13}$$

Figure 2 shows how the contours $R(ax,ay) = \varepsilon$ change with ε when both κ_{er} and κ_{ei} vanish (so $m = n = 1$ from equations (8) and (9)). At inland point $(ax,ay) = (0.715, 0.715)$, the dimensionless error $R(ax,ay)$ reaches its maximum of 8.12%. Figure 3 shows how the contours $R(ax,ay) = \varepsilon$ when $\kappa_{er} = \kappa_{ei} = 0.1a$. At inland point $(ax,ay) = (0.715, 0.7184)$, the dimensionless error $R(ax,ay)$ reaches its maximum of 8.07%. The field values of κ_{er} and κ_{ei} are $10^{-7} - 10^{-5}$ m^{-1} in estuaries or bays [*Li et al.*, 2000; *Sun*, 1997] and becomes zero in the ocean. They are always several orders of magnitude smaller than the aquifer's tidal propagation parameter a, which is usually greater than 10^{-3} m^{-1}. Hence the case $\kappa_{er} = \kappa_{ei} = 0.1a$ considered by Figure 3 shows the greatest influence of the tidal attenuation in the estuary on the groundwater level fluctuation in the L-shaped aquifer. As the inland point (ax,ay) approaches the coastlines, or infinitely far from the origin, $R(ax,ay)$ approaches zero. Given a permissible dimensionless error ε, then the approximate solution applies in areas outside the contour defined by $R(ax,ay) = \varepsilon$.

3.3. Asymptotic Solutions for Large x and y

[9] The solution's asymptotic behavior has been described by *Li et al.* [2000] qualitatively. Here it will be discussed quantitatively. From the expression of the approximate solution (11), it can be

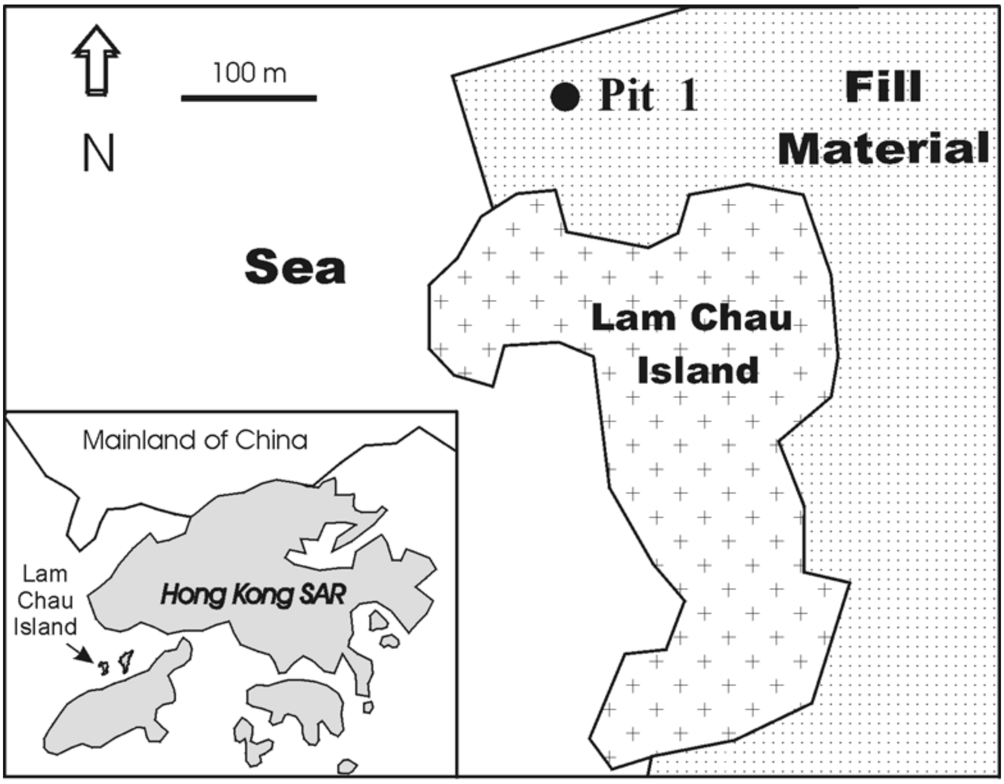

Figure 4. Location of pit 1 and Lam Chau Island in the middle of September 1994 [after *Provisional Airport Authority*, 1994].

easily seen that for large x solution (5) becomes the one-dimensional solution of *Jacob* [1950], i.e.,

$$h(x,y,t)\big|_{amx\gg1} \approx h_{\text{approx}}(x,y,t)\big|_{amx\gg1} \approx Ae^{-ay} \cdot\cos(\omega t - ay + c), \tag{14}$$

and for large y it becomes the two-dimensional solution of *Sun* [1997], i.e.,

$$h(x,y,t)\big|_{ay\gg1} \approx h_{\text{approx}}(x,y,t)\big|_{ay\gg1} \approx A\exp(-\kappa_{er}y - amx) \cdot\cos(\omega t - \kappa_{ei}y - anx + c). \tag{15}$$

Figure 3 shows that when the dimensionless distance $amx > 4.6$, the errors between (14) and (5) are within $A(0.01 + e^{-4.6})$. Figure 3 also shows that when the dimensionless distance $ay > 4.6$, the errors between (15) and (5) are within $A(0.01 + e^{-4.6}) = 0.02A$ for any $0 \le \kappa_{er} \le 0.1a$, $0 \le \kappa_{ei} \le 0.1a$. This means that when $amx > 4.6$ and $ay > 4.6$, the tidal wave interaction from the different sides of the L-shaped coastlines becomes negligible.

4. Case Study Near Lam Chou Island, Hong Kong

[10] The 12.48 km^2 Chek Lap Kok airport platform was formed in January 1996 from excavation of the original islands of Chek Lap Kok and Lam Chau and reclamation from the sea. The fill material becomes an unconfined aquifer with impermeable alluvial clay bottom. All the original data here are taken from *Provisional Airport Authority* [1994] and *Plant et al.* [1998]. The area reclaimed around Lam Chau Island in the middle of September 1994 is shown in Figure 4. A pit (see pit 1 in Figure 4) was excavated in the fill material. The distances from pit 1 to the two sides of the right angle are approximately $x_p = 80$ m and $y_p = 40$ m, respectively. The sea-land boundary near pit 1 is a right angle. Figure 5 shows the observed tidal data and their least squares fitting expressed by

$$h_{\text{tide}}(t) = h_0 + \sum_{i=1}^{2} A_i\cos(\omega_i t - c_i), \tag{16}$$

where the datum of tidal level is mPD (meters above Principle Datum) and h_0(=1.61 mPD) is the mean sea level. The seven parameters in (16) are determined by least squares fitting to the observed data. The results read $h_0 = 1.61$ mPD, $A_1 = 0.36$ m, $A_2 = 0.58$ m, $\omega_1 = 0.507$ h^{-1}, $\omega_2 = 0.237$ h^{-1}, $c_1 = 2.138$, and $c_2 = 3.209$. Equation (16) is based on the fact that the sea tides in Hong Kong are composed mainly of semidiurnal and diurnal sinusoidal components [*Hong Kong Observatory*, 1997]. The amplitudes A_1 and A_2 are much less than the average thickness $b = 8.5$ m, so the solution (5) is applicable.

[11] The circles in Figure 6 are the observed water table data at pit 1. There are 34 data, and the data from 0000 to 0930 LT on 14 September are not available. On the basis of the superposition principle and (5), the water table at pit 1 is

$$h_{\text{pit}}(t; h_p, D) = h_p + \sum_{i=1}^{2} h_i(x_p, y_p, t; D), \tag{17}$$

where $h_i(x,y,t;D)$ is defined by (5) when $\kappa_{er} = \kappa_{ei} = 0$ and the parameter a is replaced by $\sqrt{\omega_i S/(2bK)} = \sqrt{\omega_i/(2D)}$ $(i = 1, 2)$, the constant h_p is the mean water table [L] at pit 1, the constant D is the hydraulic diffusivity [L^2 T^{-1}] of the unconfined aquifer defined as $D = T/S$. In (17), there are only two unknown parameters: h_p and

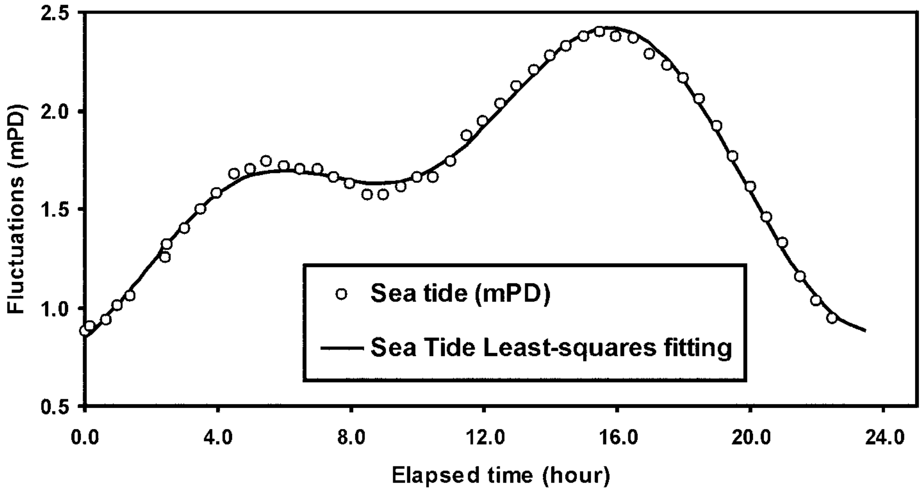

Figure 5. Observed tidal data near pit 1 from 1211 LT, 13 September 1994 and their least squares fitting expressed by equation (16).

D. All the other parameters are known. In order to estimate the two unknown parameters, the least squares problem

$$\min_{h_p, D} \sum_{j=1}^{34} \left[h_{\text{pit}}\left(t_j; h_p, D\right) - h_j^*\right]^2 \tag{18}$$

is solved, where h_j^* is the observed water table at pit 1 at time t_j (j = 1,. . ., 34). The results are listed in the second row of Table 1. The bold curve in Figure 6 presents the least squares fitting curve.

[12] By repeating the above parameter estimation procedure when the solution (5) is replaced by its approximation (11) ($\kappa_{er} = \kappa_{ei} = 0$) and Jacob's solution (14), respectively, one obtains two sets of estimated values of the hydraulic diffusivity D and the datum constant h_p. The results are listed in the third and fourth rows of Table 1, and the least squares fittings are presented by thin and dashed curves in Figure 6, respectively.

[13] It can be seen from Table 1 that the least squares residuals of solutions (5) or (11) are much less than the residual of solution (14). In Figure 6 the fitting curves of solution (5) and (11) are closer to the observed data than that of Jacob's solution. The diffusivity values estimated by solutions (5) and (11) are close to each other, but Jacob's solution overestimates the diffusivity significantly because the west boundary is ignored. The diffusivity estimated by solution (5) is 854 $m^2\ h^{-1}$, the approximate solution (11) underestimates it by only 6.8%, but the Jacob's solution (14) overestimates it by 37.5%.

5. Conclusions

[14] A periodic analytical solution is derived to improve the nonperiodic solution by *Li et al.* [2000] for L-shaped aquifers. Compared with the solution developed by *Li et al.* [2000], the

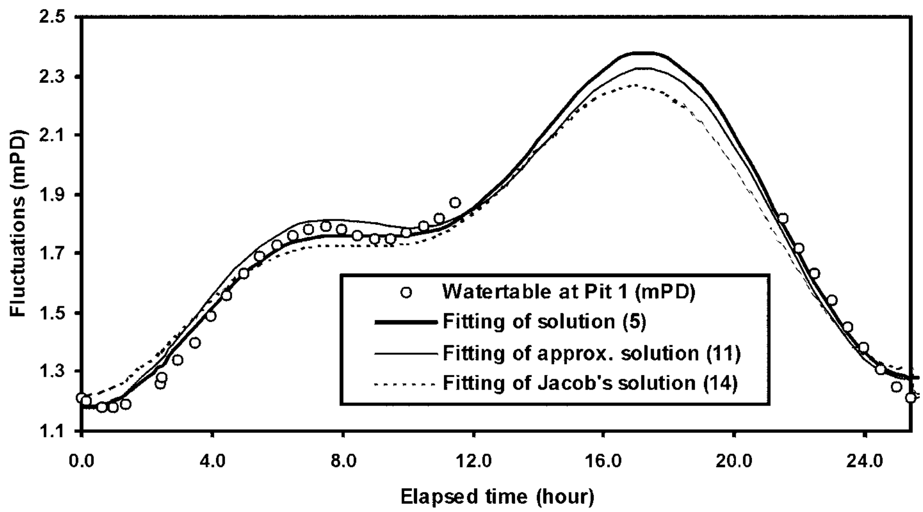

Figure 6. Observed water table fluctuations data at pit 1 from 1211 LT, 13 September 1994 and their least squares fittings based on different solutions.

Table 1. Diffusivity of Unconfined Aquifer Near Fit 1 Estimated by Least Squares Fitting Method Based on Different Solutions

	D, m^2 h^{-1}	h_p, m	Least Squares Residual, m^2
New solution (5)	854.0	1.77	0.031
Approximate solution (11)	796.0	1.79	0.018
Jacob's [1950] solution (14)	1174.0	1.74	0.126

solution here has advantages such as simplicity and periodicity. It is a solution subject to a periodic boundary condition and no initial condition is included; this is convenient for its application. Although both solutions include integrals that have to be evaluated numerically, the integral in our solution is independent of time, and hence our solution is computationally more efficient than that of *Li et al.* [2000]. Once the integral in (5) is evaluated, the solution (5) can determine the tidal groundwater level for any time; the solution of *Li et al.* [2000], however, has to reevaluate the integral for different time because the integral depends on time. A more simple approximation of the solution without integral is also presented. An error analysis and a case study show that the approximate solution has adequate accuracy in both groundwater level prediction and parameter estimation of an L-shaped aquifer. The case study shows that the solution can produce much better fitting to the observed water table and more reliable parameter estimation than *Jacob*'s [1950] solution does.

Appendix A: Derivation of the Solution

[15] Assume that

$$h(x,y,t) = A\mathrm{Re}[U(x,y)\exp(i(\omega t + c))], \tag{A1}$$

where $U(x,y)$ is a complex function, Re denotes the real part of the followed complex expression, $i = \sqrt{-1}$. Substituting (A1) into (1), (2), (3), and (4), and extending the three resultant real equations into complex ones with respect to the unknown complex function, yields

$$\frac{\partial^2 U}{\partial x^2} + \frac{\partial^2 U}{\partial y^2} = \frac{i\omega S}{T} U, \quad 0 < x, y < +\infty, \tag{A2}$$

$$U(0,y) = \exp(-\kappa_e y), \quad y > 0, \tag{A3}$$

$$U(x,0) = 1, \quad x > 0, \tag{A4}$$

$$\lim_{x\to\infty} \frac{\partial U}{\partial x}(x,y) = \lim_{y\to\infty} \frac{\partial U}{\partial y}(x,y) = 0. \tag{A5}$$

Because $e^{-(1+i)ay}$ satisfies (A2), (A4), and (A5), $e^{-\kappa_e y-(m+in)ax}$ satisfies (A2), (A3), and (A5), let

$$V(x,y) = U(x,y) - e^{-(1+i)ay} - e^{-\kappa_e y-(m+in)ax}, \tag{A6}$$

then $V(x, y)$ satisfies

$$\frac{\partial^2 V}{\partial x^2} + \frac{\partial^2 V}{\partial y^2} = 2ia^2 V, \quad 0 < x, y < +\infty, \tag{A7}$$

$$V(0,y) = -e^{-(1+i)ay}, \quad y > 0, \tag{A8}$$

$$V(x,0) = -e^{-(m+in)ax}, \quad x > 0, \tag{A9}$$

$$\lim_{x\to\infty} \frac{\partial V}{\partial x}(x,y) = \lim_{y\to\infty} \frac{\partial V}{\partial y}(x,y) = 0 \tag{A10}$$

Using the Green's function

$$G = E(x,y;x_0,y_0) - E(x,y;-x_0,y_0) + E(x,y;-x_0,-y_0) - E(x,y;x_0,-y_0), \tag{A11}$$

where $E(x, y; x_0, y_0) = (1/2\pi)\, K_0\, (\rho(a(x - x_0), a(y - y_0); 1 + i))$ [*Shimakura*, 1992], $K_n(\rho)$ denotes the modified second-kind Bessel function of nth order, implementing the standard Green's function method to solve boundary value problem (A7), (A8), (A9), and (A10) [see, e.g., *Shimakura*, 1992, p. 43], one obtains

$$V(x,y) = \int_0^\infty V(x_0,0) \left.\frac{\partial G}{\partial y_0}\right|_{y_0=0} dx_0 + \int_0^\infty V(0,y_0) \left.\frac{\partial G}{\partial x_0}\right|_{y_0=0} dx_0. \tag{A12}$$

Using the formula [e.g., *Shimakura*, 1992, p. 25]

$$\frac{dK_0}{d\rho} = -K_1(\rho), \tag{A13}$$

one finds

$$\int_0^\infty V(0,y_0) \left.\frac{\partial G}{\partial x_0}\right|_{x_0=0} dy_0 = I(ay, ax; 1+i, 1+i), \tag{A14}$$

$$\int_0^\infty V(x_0,0) \left.\frac{\partial G}{\partial y_0}\right|_{y_0=0} dx_0 = I(ay, ax; m+in, 1+i). \tag{A15}$$

Substituting (A14) and (A15) into (A12), and then (A12) into (A6), yields

$$U(x,y) = I(ax, ay; 1+i, 1+i) + I(ay, ax; m+in, 1+i) + e^{-(1+i)ay} + e^{-\kappa_e y-(m+in)ax}. \tag{A16}$$

Finally, substituting (A16) into (A1) yields the solution (5).

[16] **Acknowledgments.** This research is supported by the Committee on Research and Conference Grants (CRCG) at the University of Hong Kong. Many thanks to Li Ling and P. Nielsen for their detailed comments and suggestions.

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K. Cheung, J. J. Jiao, and H. Li, Department of Earth Sciences, University of Hong Kong, James Lee Science Building, Pokfulam Road, Hong Kong, People's Republic of China.
M. Luk, Hong Kong International Airport, 8 Chun Yueroad, Lantau, Hong Kong, People's Republic of China.