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Memo

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Date: March 15, 2014
To: John Holt, Lamoine Planning Board Chair
From: Robert Gerber, C.G.
Subject: Review of Summit Report on additional exploration at MacQuinn Pit

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I reviewed the report entitled "Supplemental Hydrogeologic Assessment" prepared by Michael Deyling of Summit Environmental Consultants in December 2013. This report was prepared in response to a Planning Board request that the applicant, Harold MacQuinn, Inc., for a gravel pit expansion of the Kittridge Pit into Lot 31 provide additional geologic information. The specifications for the acquisition of additional geologic data originally came from a report I wrote to the Planning Board on April 16, 2013, which was a peer review of the original Summit report on the geology of the site.

I believe the Planning Board actually passed a written motion that directed the applicant to do this additional work, but I don't know the exact wording of it. Therefore, I do not know if all of the things that I requested to be done were incorporated into motion.

The work and resulting report by Summit has gone a long way to answering some of the fundamental questions that bear on the potential impact of the proposed pit on Cold Spring and where the deep groundwater table lies beneath the proposed pit expansion. Before I finish my peer review of this latest report, I ask the Planning Board to consider asking the applicant for the following information to enhance the report and make it easier for me to complete my report:

1) I requested two rounds of water level data after the wells were installed. I only see one round of data summarized for PB-1, -2, and -3 in Table 1 of the report. It would be helpful to have another complete synoptic (acquired at the same time) round of water level readings. For the shallow wells, the water level readings should be taken within the next month. For the deeper wells, it is hard to tell when the "seasonal high water table" condition may be reached. I have monitored wells in deep sand and gravel and had a continuous stream gage on Libby Brook for the past 13 years in TD19 as part of my monitoring of blueberry barren irrigation for the Passmaquoddy Indians. The median peak in streamflow for Libby Brook, which drains a large glaciomarine delta, has occurred around April 1st. However, the wells, which typically penetrate 50 to 60 feet of unsaturated

sand and gravel above the water table, peak about 4 to 6 weeks later. In the MacQuinn case where the unsaturated zone is on the order of 3 times this amount, the time of the annual peak could be late summer into fall. However, the annual variation in water table is usually only a few feet, so the timing of water level measurements is not so important due to the low variability. But I still think it would be helpful to have another round of water levels taken before April 15th in all the new wells on the site plus MW-2, -3, and -4.

2) In the portion of the new report that discusses the water balance, I do not see a discussion of how the measured flows relate to any statistical measure of what those flows represent in terms of whether they are baseflows only (what was the antecedent precipitation history?) and whether these flows represent "average annual" base flows, fall high baseflows, etc. By comparison with a USGS gaged stream (Libby Brook might be similar) of similar properties and precipitation regime, one should be able to put the flow rates into some perspective.
3) For the comparison of the measured base flows with estimated flows from recharge area, it is clear that not all of the recharge area is of uniform recharge capability. I suggest dividing the recharge area into units of similar recharge capability and multiplying these sub-units by a representative recharge rate for the respective units and summing those to make the comparison. I have attached a paper that I co-authored with Dr. Charles Hebson that provides one way to do this calculation.
4) Page 7 of the PB-4 boring log is missing from the electronic file that I downloaded from the Town of Lamoine website. Can you please provide this?
5) In my recommendation for this study I specifically asked that the elevations and *locations* be surveyed with survey-grade GPS equipment. I see the elevation data attached to the new exploration points, but I saw no coordinate data. I have already spent a lot of time trying to georeference plans from the first report so that I could construct a good database in ArcGIS. I would rather not have to georeference these PDF plans to make them fit what I already have, as I did not include the time to do that in my estimate for this phase of work. Therefore, I ask for a table of x,y,z coordinates and elevations of all the new geologic explorations. As long as I know what horizontal and vertical datums are used, I can quickly add these to my database.
6) I see on Figure 1 of the new study a string of six "CSW" well locations. I have not seen drilling logs or groundwater elevations for these wells presented in either the original report or this report. I also am not aware that anyone else has entered that data into the record of this proceeding. Can this information be made available and put in the record? It would help to clarify the geologic interpretation. Were the locations of these wells surveyed by the applicant? If not, where did the applicant get the location data?
7) Michael Deyling should put his CG stamp on the report and sign it. Perhaps he did this on a cover letter or other page I do not have, but this is a standard requirement of the Board of Certification for Geologists and Soil Scientists for information provided in a regulatory proceeding.

If the Planning Board can request these clarifications then I can proceed in short order to wrap up my review of the hydrogeologic aspects of this application. If the Planning Board wants me to proceed on the basis of what data I already have, I can do that except that the margin of certainty of the meaning of the data will be less.

Attachment: Gerber and Hebson recharge reference

Geological Society of Maine

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Selected Papers on the Hydrogeology of Maine

edited by

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GROUND WATER RECHARGE RATES FOR MAINE SOILS AND BEDROCK

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ABSTRACT

This paper reviews methods and results of estimating recharge rates for Maine soils and bedrock. The percentage of precipitation that ultimately recharges soil and bedrock is an important variable in water supply safe yield evaluations and ground water contamination studies in Maine. The recharge rate is, for practical purposes, impossible to measure directly. Recharge rate varies seasonally and is related to features such as soil permeability and anisotropy, land slope, land cover, and location relative to ground water divide. Temperature variation as it affects snowmelt and evapotranspiration is also a governing factor. Recharge rates are examined for five basic geologic terrains: 1) ice contact and outwash sand and gravel; 2) glaciomarine and glaciolacustrine clay-silt; 3) thick fine-grained lodgment till; 4) thick sandy (granitic derived) till and diamicton; and 5) thin sandy widely-graded soil over bedrock. Recharge rates for bedrock overlain by these surficial materials are also examined. Historical safe yield data, baseflow as determined from stream gaging, and recharge calibrated for models of these types of aquifer systems are summarized. The results can be applied to a wide variety of ground water problems in Maine and hydrogeologically similar locales.

INTRODUCTION

The percentage of precipitation that ultimately recharges soil and bedrock is an important variable to many hydrogeological evaluations in Maine. For example, water supply safe yield evaluations and ground water contamination studies require a knowledge of net flux through the geologic formation of interest. Predictions of the highest and lowest positions of water tables also require a knowledge of response to precipitation. Recharge rate is almost impossible to measure directly, although the response of water wells as a function of precipitation is easy to measure. Unfortunately, mass balance models that compute water table rise from specific yield are overly simplistic. Even if one knows how much net recharge reaches a shallow soil water table, it is much more difficult to determine how much of that recharge will pass through to a fractured bedrock aquifer.

This paper considers several topics related to recharge rates for Maine soils and bedrock and similar hydrogeologic settings:

- a systems framework for considering recharge to ground water;
- methods for estimating net recharge to a geologic formation of interest;

- estimates by Robert G. Gerber, Inc. (RGGI), Freeport, Maine, and the U.S. Geological Survey (USGS).

From these studies, generalizations on regional average net recharge rates for different types of geologic formations are presented. In closing, further research efforts that would refine these estimates are suggested.

A SYSTEMS DESCRIPTION OF RECHARGE AND GROUND WATER FLOW

An exhaustive review of the literature on estimating recharge to ground water is beyond the scope of this article. For all practical purposes, the preferred method of direct observation and measurement is rarely feasible. Therefore, recharge must be calculated from those relevant variables that are observable or can be estimated independently. The very act of calculating recharge implies a particular model of recharge as related to other hydrogeologic variables. Modeling, whether simple or complex, plays an essential role in the estimation of recharge. The models can range from simple water balance analysis to sophisticated three-dimensional variably saturated numerical simulation. Most requirements for a realistic recharge estimate can be met by using a method intermediate in accuracy, complexity, and cost.

Given that some kind of modeling is generally required to estimate recharge, it is helpful to establish the systems nature of ground water flow. Precipitation recharge is usually the single most important natural input to a hydrogeologic system, while discharge to surface water is the ultimate system output. Piezometric head describes the state or condition of the system. The system parameters (hydraulic conductivity, storativity, fracture characteristics) and configuration (topography, stratification, bedding, types of rocks and soils) are fundamentally geologic. Important external forcing functions include pumping and artificial recharge, climate, water extraction by vegetative evapotranspiration (ET), and connections to adjacent aquifers and surface water bodies. There is essentially one physical process of interest, porous media flow as described by Darcy's law in saturated and variably saturated systems. The entire system must obey fundamental mass balance constraints.

Even though recharge is formally a system input in this context, it also is frequently referred to as a model parameter. This semantic distinction, while real, need not be an obstacle in this discussion. Even though recharge is an input, it also is unknown, and therefore must be estimated. Following conventional usage, recharge estimation will be referred to as a problem of parameter estimation in this paper.

The important factors affecting recharge can now be given within the systems context described above. Recharge rate varies seasonally in Maine and is complexly related to landscape, climatic, and hydrogeologic characteristics, including:

- hydraulic conductivity and anisotropy
- specific yield
- unsaturated flow behavior
- topography: slope and relief
- surface location relative to ground water divide

- soil thickness
- vegetative land cover type
- precipitation distribution in time and space
- temperature as it affects snowmelt and ET

METHODS FOR ESTIMATING RECHARGE

The extent to which all of the above factors are included in an analysis, i.e. the extent to which the system is modeled, will dictate the complexity of the method for estimating recharge. Four general approaches to estimating recharge are considered here:

- traditional water balance analysis
- parameter estimation from observations on measurable quantities,(e.g. ground water levels and stream flow), in conjunction with ground water flow models
- direct estimation based on conceptual and/or physically-based models of hydrologic processes
- indirect estimation by saturated and variably saturated flow modeling

Water balance analysis as applied to ground water systems is the simplest approach. Beyond that, it is difficult to assign a scale of difficulty to the other three approaches. Within each category, there is a range of complexity from which to choose.

The principle of mass balance is fundamental to all of these approaches. Water balance analysis is nothing more than hydrologic bookkeeping. Both automatic and trial-and-error paramèter estimation may use numerical methods, statistics, and mathematical optimization in conjunction with flow modeling. Direct estimation by modeling the component hydrometeorological and vadose zone processes is a mechanistic approach that models the hydrologic cycle from initial precipitation down to percolation (recharge) at the water table. Indirect estimation uses fully saturated or variably saturated flow models with boundary conditions such that the models implicitly calculate recharge to the aquifer. Variably saturated flow modeling avoids the need for explicitly identifying ground water recharge rates, but does require estimates of the various hydrologic abstractions at the surface and in the unsaturated zone.

Water Balance Analysis

Water balance analysis is based on applying the principle of mass conservation to the ground water system of interest. The aquifer is treated as a lumped system, with fluxes integrated over the system. All fluxes besides recharge must somehow be estimated, with recharge to ground water given as the remainder in the mass balance equation. In essence, it is a form of bookkeeping applied to the various fluxes in the hydrologic system.

Water balance methods have several advantages:

- (potential) simplicity
- conceptually appealing

- physically consistent
- a long history of application in hydrology

Water balance analysis, though well-suited for application over large systems at steady state, is not without some limitations:

- estimation of the component fluxes can sometimes be a problem, especially evapotranspiration (ET)
- water balance analysis may be too coarse for smaller systems where a higher degree of resolution is required

The water balance equation can be written as

$$P = RO + R + ET$$

where P = total precipitation, RO = runoff, R = recharge, and ET = evapotranspiration. Given measurements for P and independent estimates of RO and ET, R can be calculated.

Runoff is simply understood to be the fraction of precipitation that runs off the land and is not available for infiltration and eventual recharge. The best estimates of runoff are derived from actual stream gaging and precipitation data. In the absence of data, runoff is frequently calculated in practice with the Soil Conservation Service (SCS) Runoff Curve Number (RCN) method (Mockus, 1964). The RCN is a function of soil type, soil drainage condition, and land cover type. Evapotranspiration can be estimated with, among other techniques, one of the Penman vapor diffusion/energy balance methods (e.g., Eagleson, 1970, pp. 211-242), a site-specific empirical method, or Morton's (1983) complementary relationship. Johnson (1977) concluded that the hybrid "Goldschmidt-Thornthwaite" theory was most appropriate for temperate continental areas such as New York state. On a watershed-wide basis, averaged over a year, analysis of the U.S. Geological Survey stream gaging data from Maine shows that 50% to 65% of precipitation passes through the watersheds. By simple water balance calculation, therefore, the average net ET in Maine is in the range of 35% to 50% of precipitation. Complex empirical formulas are unlikely to estimate the ET component much more accurately over an average year. However the formulas should provide insight as to how the ET can vary as a function of season and unique site-specific conditions. A limited amount of spatial and temporal distribution of fluxes can be handled by applying mass balance over smaller subareas. However, if local resolution, spatial variation, or seasonal effects are of primary interest then some of the other methods discussed here may be more appropriate. Alternatively, more detailed watershed models might be considered (e.g., Leavesley and others, 1983; Famiglietti and others, 1992). This kind of detail is probably most important in shallow, unconfined aquifers; flow in deeper bedrock aquifers is likely less affected by flux variability at the surface.

Parameter Estimation and Inverse Methods

The problem of estimating aquifer properties from water level observations has been called the "inverse problem" of ground water modeling. Formal parameter estimation utilizes sophisticated methods of optimization, statistics, and systems theory in conjunction with numerical models of ground water flow. These methods are called "inverse methods" or "automatic calibration". Yeh (1986) gives a review of parameter estimation in ground water hydrology. Model parameters (e.g., transmissivity, recharge) are estimated such that a specified error criterion (e.g., sum of squared deviations between observations and simulation) is minimized. Prior to the development of formal algorithms for parameter estimation, models were typically calibrated by systematic trial-and-error (as opposed to "automatic") adjustment of parameters. The goal of manual calibration is also one of minimization, often the root mean square error (rmse) or the mean absolute error calculated from differences between simulated and measured water levels. Calibration and recharge estimation are improved if streamflow gain/loss data are also available, in addition to water level data. Inverse methods are potentially very powerful and in the right hands may be preferable to other methods for recharge estimation in typical Maine applications.

Advantages of inverse methods for estimation of recharge include:

- estimates are consistent with underlying conceptual and mathematical flow models
- temporal and regional spatial variability can be accommodated
- estimates are optimal according to a defined criterion
- knowledge of ET and soil water distribution not required

Despite these strong points, automatic inverse methods are not yet routinely used. This may be attributable to some of the following disadvantages:

- high level of analytical and computer expertise required
- physically unreasonable estimates possible (may indicate problem in conceptual model)
- optimal estimates may be non-unique
- independent estimates of transmissivity may be needed
- unknown parameters may not all be identifiable and observable
- calibrate only to heads or fluxes; unable to calibrate to specified gradients

The related problems of identifiabilty and observability bear special mention. In a prototype vertical cross-sectional model bounded by a divide (no-flow boundary) and a discharge point (fixed head boundary), aquifer parameters governing steady-state saturated flow are recharge and transmissivity. The flux through the aquifer and the predicted water table positions will be a function of the ratio of recharge and transmissivity. A reasonable match between predicted and observed head does not guarantee a unique recharge estimate, since it is the parameter ratio that is critical. Knowing in advance the approximate recharge rate narrows the possible range of transmissivities. These issues lead directly to the subtleties of singularity, identifiabilty, and observability in parameter estimation (e.g., Cooley and Naff, 1990, pp. 77-81).

Several parameter estimation computer programs are generally available, in addition to the larger number of research codes with limited circulation. Cooley and Naff (1990) of the USGS have developed a parameter estimation computer program based on linear and nonlinear regression in combination with a two-dimensional integrated finite difference ground water flow model. Doherty (1990) has developed a regression program, MODINV, for use with the USGS MODFLOW three-dimensional flow model (McDonald and Harbaugh, 1988). Hill (1992) recently published the MODFLOWP regression module for use with MODFLOW. MODFLOWP is quite flexible and can be configured to a wide variety of estimation problems, including recharge estimation. Despite the availability of these codes, trial-and-error calibration will continue to be widely used, especially in simpler applications.

One-Dimensional Direct Modeling of Recharge

Direct modeling uses a decoupled model of hydrologic processes in the soil column above the water table. Most of these models have come out of agricultural engineering and soils science, where the object is to model water and chemical fluxes through the unsaturated zone. However, it is just a small conceptual leap from flux through bottom of the column to ground water recharge. With modification, these models can be used to calculate recharge rates to ground water. These models are typically one-dimensional in the vertical. The major component physical processes are included (e.g., ET, actual water evaporation and consumption by plants, interception of precipitation, infiltration, runoff, and redistribution). Thus, these models may be thought of as detailed, physically-based water balance models. The treatment of the component submodels may be conceptual or physically-based, the latter usually at the cost of added complexity and data requirements. The models are decoupled from the ground water regime. The lower boundary condition may be set as the water table in the case of shallow systems or as a specified head gradient (commonly = 1) for steady gravity drainage at depth. These models are data-intensive and require continuous and/or event meteorologic data, detailed soils characterization, and crop/plant growth and water consumption parameters.

A significant limitation of these recharge models is that they are one-dimensional. Variables such as distances to ground water divides and land slope are not considered. It is intuitive that steeper land slopes will shed more water and allow less deep percolation than flat land, other things being equal. This effect is not quantified in any meaningful way in the one-dimensional vertical models. However, the form of the landscape and the relationship of points to the ground water divides and discharge areas can be taken into account where water tables are close to ground surface. A useful technique for analyzing this special case is presented in a later section of this paper.

Advantages of direct recharge calculation by vertical modeling can be summarized as:

- includes detailed temporal variability due to plant growth and climate
- reasonable computational burden, though a computer is required

Corresponding disadvantages are:

- moderately heavy data requirements
- cannot capture spatial effects of topography and location between divide and discharge boundaries

- moderate modeling expertise required

A number of fully-documented, public domain one-dimensional models of vertical flow above the water table are available:

- **CREAMS**
- **GLEAMS**
- **HELP**
- **DPM**

All of these models use simplified soil water routing to simulate vertical flow through the unsaturated zone. Such an approach works best in humid climates.

CREAMS (Knisel, 1980) and its extension, **GLEAMS** (Leonard and others, 1986) are perhaps the best known and most widely used models in this list. The name "**CREAMS**" is an acronym for "**C**hemicals, **R**unoff, and **E**rosion from **A**gricultural **M**anagement **S**ystems". The model uses SCS curve number hydrology for runoff calculations, a simple soil water storage routing model for vadose zone water movement, and Ritchie's (1972) potential evapotranspiration model. A crop growth/water demand component is also part of **CREAMS**. **CREAMS** was later extended to include aspects of hydraulic and chemical loading to ground water. The result was dubbed **GLEAMS**, for "**G**roundwater **L**oading **E**ffects of **A**gricultural **M**anagement **S**ystems". However, it has been reported that **GLEAMS** does not route water completely from the root zone to the water table (Shoemaker et al., 1990).

The EPA-sponsored **HELP** model (Schroeder and others, 1984a,b) is another offshoot of **CREAMS**. **HELP** (**H**ydrologic **E**valuation of **L**andfill **P**erformance) was developed by the U.S. Army Corps of Engineers for EPA using **CREAMS** as a starting point. HELP is intended for calculating leachate fluxes through landfill caps and liners. It is often used to calculate recharge to ground water beneath landfills. One of the major changes from **CREAMS** was the inclusion of modules for calculating flows through engineered landfill drains. Most of the component submodels are similar to **CREAMS**. Since **HELP** is designed specifically for landfill analysis and design, it is of limited use for estimating regional recharge.

The last model in this group is the U.S. Geological Survey **D**eep **P**ercolation **M**odel, or DPM (Bauer and Vaccaro, 1987). The model is conceptually similar to **CREAMS**, **GLEAMS**, and **HELP** in terms of the physical process that are included and the manner in which flow through the column is modeled. Just as **HELP** is specialized for landfill applications, **DPM** is specialized for calculating deep percolation rates. Therefore, of these four models it may be preferred for estimating regional recharge.

There may be occasions in Maine when the unsaturated zone is thick, or the unsaturated zone requires a more physically based treatment. Then a one-dimensional variably saturated flow model should be considered, e.g., **Opus** (Smith, 1992) or **RZWQM** (GPSR, 1992). Some of the multi-dimensional variably saturated flow models listed below can also be applied to vertical one-dimensional situations.

Indirect Estimation of Recharge: Saturated and Variably Saturated Flow Modeling

In both water balance analysis and one-dimensional vertical modeling, recharge (or vertical flux through the bottom of the soil column) appears as an explicit boundary flux. The models, however simple or complex, solve for the recharge quantity when all other terms are known. In contrast, recharge does not appear explicitly in indirect methods, but is calculated as an internal flux in the model. This usually means that some effort may be required in interpreting model results to extract the desired recharge estimates. The advantage of this approach is that the problem can be posed in terms of known information (e.g. water levels) and system boundary conditions. Provided the information is accurate and the model is conceptually correct, the resulting recharge estimates should be consistent with the known information.

Advantages of estimating recharge indirectly with a flow model include:

- problem is posed in terms of known and observable boundary conditions
- recharge is calculated implicitly
- recharge estimate is consistent with underlying conceptual model
- allows for spatial distribution of recharge

The disadvantages of this approach are closely related to the advantages listed above:

- boundary conditions must be known
- conceptual model must be consistent with the actual hydrogeology
- modeling expertise and supporting computer resources are required

Saturated Flow Modeling

Indirect estimation of recharge with saturated flow models can be approached with two-dimensional models in section or plan, or three-dimensional models. It is appropriate for steady or transient flow situations. The essence of this approach is to develop the flow model in terms of known third-type (Cauchy or "leaky") boundaries over part or all of the domain. This boundary condition is formally given in terms of hydraulic head in a "leaky adjacent aquifer" separated from the aquifer of interest by an aquitard. In practice, though, this method can be used even when a clearly defined aquitard is not present. The primary requirements are that the phreatic (or piezometric) surface and hydraulic conductivity are known or can be estimated over the entire domain.

In section models, the water table position (or reference head surface, for leaky connected aquifers or water bodies) is assumed known, and the goal is usually to determine the distribution of recharge across different strata in the aquifer of interest. Section models are quite useful in that the spatial distribution of recharge (and discharge) along the section is determined. This approach is best used at a local or small regional scale.

In plan models, recharge across a semi-confining layer into a lower unit is often of interest. The same kind of known information assumed in section models is also assumed in plan models. Two- or three-dimensional models can be used, though three-dimensional models are most representative of the physical flow system. As saturated flow modeling is now a mature subject that is widely applied, it is not necessary to tabulate the many available models. In passing, though, the popular three-dimensional finite-difference USGS **MODFLOW** model (McDonald and Harbaugh, 1988) and the flexible two-dimensional finite-element **AQUIFEM** model (Townley and Wilson, 1980) are considered by the authors to be among the best.

Variably Saturated Flow Modeling

The use of variably saturated flow models enables one to avoid the problems of specifying ground water recharge rates or calculating recharge as distinct from the ground water system. Instead, the effective recharge at the ground surface is specified. The model then determines the movement of water in the unsaturated and saturated zones, including some or all of the aquifer of interest. This has the advantage that the internal spatial distribution of flux across the water table (i.e. recharge) is automatically determined by the model. Variably saturated flow models may be most useful when the temporal distribution of recharge is required, the unsaturated zone is of significant thickness, and there is significant ET demand in the shallow (root) zone. Thick unsaturated zones often induce long lag times between application of precipitation at the surface and recharge at the water table. This effect cannot be captured by saturated flow models.

Much of the original work in variably saturated flow modeling was motivated by agricultural and soil physics needs, but more recently the unsaturated zone has received attention in studies of potential nuclear waste repositories and landfill failure analysis. Since much of the original work in unsaturated flow originated in agricultural work, many models include well-developed algorithms for including plants and evapotranspiration as moisture sinks. However, the effective precipitation at the ground surface must still be estimated, considering the surface processes of interception, runoff, infiltration, etc.

As with fully saturated section models, multi-dimensional variably saturated flow models produce a physically-based spatial distribution of recharge as a function of geology and topography. This is useful if a detailed understanding of recharge/discharge patterns is required. Inverse methods applied to regional flow models and one-dimensional recharge modeling cannot produce this level of detail.

A representative list of available public domain two-dimensional models follows:

- VS2D (Lappala et al., 1987) and VS2DT (Healy, 1990)
- UNSAT2 (Davis and Neuman, 1983)
- SUTRA (Voss, 1984)

Fully three-dimensional models are impractical for most problems, due to computational and parameter demands. The nature of unsaturated flow is such that even the one- and two-dimensional models should only be used by those experienced in the computational and theoretical aspects of unsaturated flow.

Choosing a Method

The most useful and general recommendation is to consider only those methods that address the important features in the problem under consideration. Anything less, and one is likely to obtain an inadequate model for estimating recharge. Anything more, and one is likely to expend scarce resources on a needlessly complex model. Furthermore, higher levels of complexity generally imply greater data requirements. It is difficult to justify a particular method if the supporting data is not available. For example, a transient three-dimensional variably saturated flow model is inappropriate if average annual regional recharge to a sand and gravel aquifer is of interest.

It is good practice to include a simple water balance analysis in any determination of recharge, even if more complicated methods are used. Water balance provides a check on the realism of results obtained by other methods. Part of basic water balance analysis is a determination of ET. Several methods for estimating ET have already been mentioned in the "Water Balance Analysis" section. ET calculations are subject to great error, and care must be exercised when interpreting the results of whichever method is used.

On a regional basis over large time scales the water balance method is probably sufficient. At smaller scales or for transient analysis, one of the other methods may be necessary. The other methods all have the drawback that they require higher degrees of sophistication and expertise on the part of the hydrologist. They also require significantly more data and computing resources. Confident, knowledgeable use of the more advanced methods presupposes experience as well as advanced study in theoretical and computational hydrology, as none of the methods are trouble-free. At a minimum, a basic understanding of the physical processes and hydrogeologic setting are required.

If a numerical ground water model is being applied, then automated inverse methods should be considered. They constitute a systematic approach to the difficult task of calibrating numerical flow models, and are useful for both transient and steady-state applications. The resultant recharge estimate is inherently consistent with the rest of the ground water model as well as the underlying conceptual model. Thus, automated inverse methods are also useful for evaluating the soundness of the conceptual model. Models so calibrated should be checked against hydrogeologic experience, suggested "rule of thumb" parameter values, and an independent water balance analysis. One weakness of currently available codes is that they only calibrate to heads or fluxes, and not specifically to horizontal or vertical gradients. Critical gradients should be checked carefully in models calibrated with automated methods. Recharge estimates from manual or automatic inverse methods are generally better if good independent estimates of transmissivity are available. Then transmissivity can be taken as "known" in the inverse modeling. There is usually more difficulty in estimating transmissivity than in estimating recharge rates. This is because transmissivity can easily vary over an order of magnitude while recharge is more tightly constrained. If automated methods are inappropri- ate, then trial-and-error can be used to estimate recharge by matching simulated heads and fluxes to available data. Examples are given later in this paper of trial-and-error and automated inverse methods for recharge estimation where the hydraulic conductivity distribution was known with some degree of confidence.

An alternative to inverse modeling is to use one of the direct recharge calculation models. For typical ground water applications in Maine, DPM is probably the best choice since it was developed for ground water applications and the vertical flow component has been shown to be adequate for humid climates (Thompson and Tyler, 1984). If the unsaturated zone processes are of special concern, if there is a specific agricultural aspect of the problem, or transient behavior at the field scale is a problem, then one-dimensional variably saturated flow models might be considered. Multi-dimensional variably saturated flow models such as VS2D and UNSAT2 can also be used for one-dimensional applications. Variably saturated flow models are most useful when strong seasonal recharge variation (e.g., due to distinct growing seasons) or transient flow is an important feature, or when the unsaturated zone is thick. Their use is questionable when these factors are unimportant.

At local or small regional scales, indirect estimation with flow models may be useful for determining recharge. The chief advantage of multi-dimensional variably saturated flow models is that they produce a spatial distribution of recharge that is governed by the geology and topography in an integrated, physically based manner. When used in cross-section, these models are limited by their two-dimensional nature to a single flow path with no horizontal extent. Depending on the problem this may be acceptable. Along with automated inverse methods, these models may also be the most intimidating from theoretical, computational, and data requirement perspectives. Many of the variably saturated flow models include sophisticated treatments of plants as moisture sinks, especially useful for modeling seasonal behavior.

If flow in the unsaturated zone is unimportant, indirect estimation using saturated flow models should be considered. They are considerably easier to use than variably saturated models and yield much of the same information. Whereas variably saturated flow models only make sense in transient and seasonal applications, fully saturated flow models using leaky boundaries can be used effectively under steady or transient conditions. Transient behavior due to seasonal variation can often be modeled in a quasi-steady manner (i.e., model each season as a steady state, but vary the boundary conditions by season).

WATER TABLE RESPONSE TO RECHARGE

Before presenting ground water recharge rates for specific geologic terrains characteristic of Maine, it is useful to consider the general issue of water table response to recharge. The very notion of response implies some kind of transient behavior, as opposed to steady state conditions in which a long-term average water table position is of interest. Three problems commonly appear when interpreting the response of a phreatic aquifer to recharge.

Short-Term and Long-Term Specific Yield

A common and simplistic way of calculating the response of a water table to an increment of recharge is to divide the amount of recharge (units of length, [L]) by the specific yield. For example, one might assume that 0.25 inches of recharge on a phreatic aquifer with 0.25 specific yield will cause a 1 inch rise in the water table. Specific yield is a time-dependent variable and it is usually not the same for a drying condition as it is for a wetting condition during the transient state. A discussion of these aspects of unsaturated flow can be found in many textbooks, such as Bouwer (1978, pp. 29-31).

Pumping test analyses of wells in unconfined stratified sand and gravel aquifers will often produce a calculated specific yield of about 0.05, although the commonly accepted textbook value for these materials would be about 0.25. Over a typical year on Maine sand and gravel aquifers, with alternating wetting and drying cycles, an average value of specific yield of 0.1 seems to provide the best calibrated fit to regional models. Here the times between major change in volumetric-water content of the unsaturated zone and water table changes is longer than the duration of a typical pumping test, but shorter than the time required for complete drainage to a new "equilibrium" condition in the volumetric-water content of the unsaturated zone above the water table.

Lag Time between Recharge Application and Water Table Response

The next problem in predicting water table change from a given recharge distribution relates to the time between application of recharge and the response of the water table change. This only becomes an issue in transient evaluations. In the case of simple lumped-parameter models, the response time can be calculated explicitly. Sangrey et al. (1984) suggest that water table response is a linear function of recharge on the basis of empirical evidence. The parameters of the relation can be estimated by standard linear regression. This approach is somewhat simplistic, and due to its empirical basis the parameters lack meaningful physical interpretation.

Gelhar and Wilson (1974) used a simple water balance/linear reservoir conceptual model to study aquifer response to recharge. This same approach is used extensively in surface hydrology rainfall-runoff modeling (e.g. Dooge, 1973). In Gelhar and Wilson's application, the reservoir time constant (i.e. response time) is $T_c = SL^2/3T$, where S is specific yield (dimensionless), L is the distance [units of length] from ground water divide to discharge zone, and T [units of length^2/time] is the effective aquifer transmissivity. The response time T_c is the time required for the aquifer to complete 63% of the total change in storage induced by a step change in recharge. The value 63% is a consequence of the assumed linear reservoir conceptual model and the resultant exponential impulse response function. The time t^* to reach an arbitrary degree of completion c^* is $t^* = \{-T_c \ln(1-c^*)\}$, $(0<c^*<1)$, so that time to 95% completion is $t_{95} = 3T_c = SL^2/T$.

While these simple models may not be suitable for local site-specific applications, they are useful for developing a conceptual understanding of aquifer dynamics. For example, the response time of very large aquifers (such as the Raritan Formation in New Jersey) is much longer than those of small isolated New England glacial deposits. Most New England aquifers respond to seasonal changes in recharge pattern. Some are even small enough to respond to daily changes. However, large aquifers only respond to long-term geologic, climatic, and man- induced stresses.

Movement of Recharge Through the Unsaturated Zone

A final problem in evaluating the response of the water table to recharge is calculating the time for precipitation recharge to move through an unsaturated zone. This is different from the response time discussed above which relates solely to specific yield, transmissivity and aquifer length scale, and not to unsaturated thickness above the aquifer. The time required for flow through the unsaturated zone is small for most Maine aquifers. However, there are places in Maine, such as the sand and gravel aquifer near the McKin Chemical Superfund site in East Gray, where as much as 100 feet of unsaturated material overlies the aquifer. In order to simulate properly the response of the water table to recharge, a one-dimensional unsaturated flow model must be coupled to the saturated flow model. This had to be done, for example, in developing ground water management models for the Long Island, New York, aquifer (G. Pinder, personal communication., 1986).

RECHARGE RATES IN MAINE SAND AND GRAVEL

For sand and gravel aquifers, RGGI has identified a narrow range of recharge estimates based on results from modeling studies in Maine. Recharge values also exist for this soil type from Long Island, New York, and Cape Cod, Massachusetts, although recharge rates are somewhat less (16 to 18 inches per year according to Wilson, 1981) than for Maine. This difference is due to the higher average annual temperatures and therefore greater ET rates. Average annual precipitation in these hydrogeologically similar locales is on the order of 45 inches per year.

Surface water runoff appears to be insignificant on sand and gravel aquifers. In southern and mid-coastal Maine, stream gage analysis indicates that ET accounts for about 35% to 40% of average annual precipitation loss. It stands to reason, therefore, that in sand and gravel aquifers with no runoff that the remainder of the precipitation is ground water recharge. Cervione et al. (1972) found that only 5% of total runoff from precipitation on Connecticut stratified drift aquifers is surface runoff; the remainder is ground water discharge. Vechiolli and Miller (1973) concluded that surface water runoff was insignificant in the Ramapo River Valley aquifer, a stratified-drift valley fill aquifer in northern New Jersey.

Recharge Estimates from Ground Water Models by Robert G. Gerber, Inc.

RGGI (1984) calibrated a two-dimensional finite-element flow model of the outwash sands of the Branch Brook aquifer in Kennebunk and Wells to base flow of the brook as measured at the Kennebunk, Kennebunkport & Wells Water District (KKWWD) pumping station. The best fit was obtained with an average annual recharge rate of 55% of precipitation.

Similar recharge estimates resulted when RGGI (1987a) evaluated the long-term safe yield of the Estabrook and Stevens well field in a well-defined sand and gravel aquifer in North Yarmouth. This well field has a proven safe yield of 475 gpm. RGGI calibrated a one-dimensional model (Wilson, 1981) to this aquifer. When the pumping rate exceeds the safe yield, static water levels decline. Based on areal mapping of contributing area, the average annual recharge rate must be on the order of 60% to 65% of precipitation to produce this yield.

Recharge Estimates from Ground Water Models by U.S. Geological Survey

Computer modeling studies by the U.S. Geological Survey have developed estimates of recharge rates on sand and gravel in valley-fill stratified drift and outwash glacial aquifers. Morrisey's (1983) study of the Little Androscoggin River valley gave a recharge estimate of 45% of precipitation in water year 1981. Undoubtedly there were other soil types and urbanized areas within this large study area that would tend to reduce the recharge rate estimate from a more representative range of 50% to 55% for pure sand and gravel.

An important point in the Morrissey (1983) study is the necessity of adding the recharge contribution from adjacent till-covered uplands to the edge of the stratified drift. Based on flow model calibration, he suggests that one can estimate lateral inflow across a given edge length of sand and gravel aquifer by taking 60% of the precipitation falling on the surface watershed above the edge of the sand and gravel aquifer. This lateral influx takes the form of ground water flow as well as surface runoff that subsequently infiltrates the sand and gravel.

Tepper et al. (1990) developed a model of the Saco River Valley glacial aquifer between Bartlett, New Hampshire and Fryeburg, Maine. They cite a recharge estimate of 24 inches per year for stratified drift aquifers in the glaciated northeast United States, or approximately half of total annual precipitation. They accepted this as a "known" value and did not calibrate their model by adjusting recharge. Average annual precipitation in the study area is approximately 44 inches per year.

Recharge estimates from studies of other glaciated terrains in the northeast United States can also be used to make inferences for Maine sand and gravels. Vecchioli and Miller (1973) estimated long-term average annual recharge to the 4.5 square mile Ramapo River valley-fill stratified drift aquifer in northern New Jersey as 25 inches per year, with average annual precipitation of 45 inches per year. This estimate was based on water balance and assumed no significant surface runoff. Getzen (1977) cites a gross estimate of 23 inches per year recharge for Long Island, New York, as given by Cohen et al. (1968). He suggested that it is probably too high for the entire island, as it ignores several important factors, including spatial differences in precipitation, the distribution of morainal and outwash soils, thickness of the unsaturated zone, and local relief. He therefore developed a spatial distribution of average annual recharge for his analog flow model of Long Island, with recharge ranging from 16.9 inches per year to 21.7 inches per year. Average annual precipitation on Long Island is on the order of 45 inches per year.

In a later modeling study of the Ramapo aquifer, Hill et al. (1992) calibrated a two- dimensional **MODFLOW** application by trial-and-error adjustment of recharge and several other parameters. They obtained acceptable fits to observed water levels and measured streamflow gains and losses with a range of recharge values, from 8.8 inches per year (0.002 feet per day) to 19 inches per year (0.0042 feet per day), with a recommended compromise value of 12.4 inches per year (0.0032 feet per day). This apparently low value is due to the fact that they calibrated to seasonal low water level conditions on October 13, 1982, and is therefore not representative of long-term average annual recharge. This highlights the importance of using recharge estimates that are consistent with the available calibration data. If the goal is to estimate long-term average annual recharge, then the calibration data must reflect long-term average conditions, and not seasonal and/or other cyclical lows or highs.

Seasonal Variation of Recharge and Drought

Although average annual recharge values are used for long-term contaminant transport studies (where the time scale is on the order of tens of years or more), seasonally variable recharge values may be more appropriate for some studies. As part of the Branch Brook watershed ground water model (RGGI, 1984), a detailed water balance was performed on data recorded during a one-year test period (10/81 to 10/82). The available data included monthly water levels from USGS-installed monitoring wells within the model area and partial stream gaging of the Branch Brook watershed (D'Amore, 1983). Average monthly rates derived from the RGGI (1984) Branch Brook study are summarized in Figure 1 for sand and gravel in south-coastal Maine. The range of values reported above for Hill et al. (1992) is consistent with the recommended September and October recharge values for Southern Maine in Figure 1.

Note the net removal of water indicated by Figure 1 in June, July, and August. This would only apply on those phreatic aquifers where the water table is close enough to ground surface to be affected by ET. The removal of water by ET is included in most ground water models, e.g., McDonald and Harbaugh (1988), Prickett and Lonnquist (1971), and Trescott et al. (1976). It is typically treated as a truncated linear depth function, such that ET is a maximum when the water table is at the ground surface, and it decreases linearly to zero with the water table at a defined depth below ground surface below which ET ceases to function. The values in Figure 1 were derived from the trial-and-error inverse procedure of calibrating the Branch Brook aquifer model (RGGI, 1984) to measured water table fluctuations in a number of USGS monitoring wells spaced throughout the aquifer. Actual precipitation and calculated potential ET from measured meteorologic conditions were used to guide the calibration. The water table in much of the upper section of the aquifer is close to ground surface.

The Branch Brook study also yielded insight into the effect of drought on ground water levels and discharge. The 1965 drought in the Branch Brook aquifer was simulated such that the minimum monthly flow in Branch Brook matched the recorded yield at the KKWWD pumping station. The estimated monthly recharge and ET values for that year are somewhat site-specific and also depend upon the choice of specific yield. Subject to these qualifications, it was observed that in severe drought years precipitation is only about 60% of average annual precipitation. It is reasonable, therefore, to simulate a severe drought by multiplying average recharge rates by 60%. Surface water reference head values at first- and third-type boundaries should also be adjusted for drought.

Figure 1 includes some recharge during the winter months of December, January and February. During this period most precipitation is in the form of snow. The snow does not actually recharge the aquifer until March and April when the recharge rates are much greater due to snow melt. Recharge will occur during winter when precipitation is in the form of rain or when average daily temperature is above freezing under forested watersheds (Gerber, 1978). The distribution of recharge throughout the year will shift as one goes from southern to northern Maine. As one goes further north, relatively more water is stored during the winter months in the snow pack. This snow pack water creates a larger spring recharge event that begins later and lasts longer.

RECHARGE RATES IN SANDY GLACIAL TILL

Sandy glacial till and diamicton is located on and in the down-glacier direction from granitic rock masses in Maine. The till commonly has from 15% to 25% passing a #200 sieve (in the silt and clay range) and has a typical hydraulic conductivity of 0.15 to 1.5 feet per day (ft/day). Morrissey (1983) used a hydrograph separation method to evaluate the baseflow from the Little Androscoggin River above South Paris. The watershed is covered primarily by a sandy till derived from granitic rock. Stream baseflow varies seasonally. When an "average" baseflow figure is cited in the literature, it is an average of the values across the seasons. Morrissey (1983) found that for water year 1981, which had 39.4 inches of precipitation at West Paris, the till produced an average baseflow of 19% of precipitation or 7.4 inches. This is interpreted as the ground water recharge rate of the sandy glacial till.

During the period 1987 to 1989, a major geologic and hydrogeologic investigation was conducted in Township 30, Washington County, for a special waste landfill. This study generated useful data pertaining to recharge in glacial tills (RGGI et al., 1988). The site is situated on a granitic pluton, but metasedimentary rocks are located about 2 miles up-glacier. The fines content of the till and diamicton on the site was quite variable and ranged from 15% to 55%. Because of intermixing of some metasedimentary-derived sediments in the till matrix, the geometric mean insitu horizontal hydraulic conductivity is about 0.1 feet per day in the upper 20 feet and 0.04 feet per day below 20 feet depth, which is slightly lower than a till derived from only granitic rock. However, hydraulic conductivity varied over a wide range and a uniform horizontal hydraulic conductivity of 0.2 feet per day was used to represent the till in the final model.

A regional three-dimensional model was constructed using **MODFLOW** (McDonald and Harbaugh, 1988). This model was calibrated to 38 monitoring wells, including numerous clusters of wells measuring potentiometric levels at various depths. Forty-three in-situ permeability tests and 3 bedrock pumping tests provided information on hydraulic conductivity. The model was calibrated with a recharge rate of 5 inches per year applied over an area where till was prominent on the surface and 5.5 inches per year where ice disintegration deposits dominated. The recharge was applied uniformly over both recharge and discharge areas.

The model calibration, although good (arithmetic mean of residuals of 0.9 feet and mean absolute value of residuals of 4.9 feet), could have been improved by spatially distributing recharge while maintaining an overall average for the model area of 5 or 5.5 inches per year. Nonetheless, most modeling studies use a uniform recharge because of the difficulties of determining the spatial distribution of recharge.

Sensitivity analyses on the till and ice disintegration recharge rate using **MODFLOW** showed that multiplying the calibrated recharge by a factor of 1.5 caused an average poten-

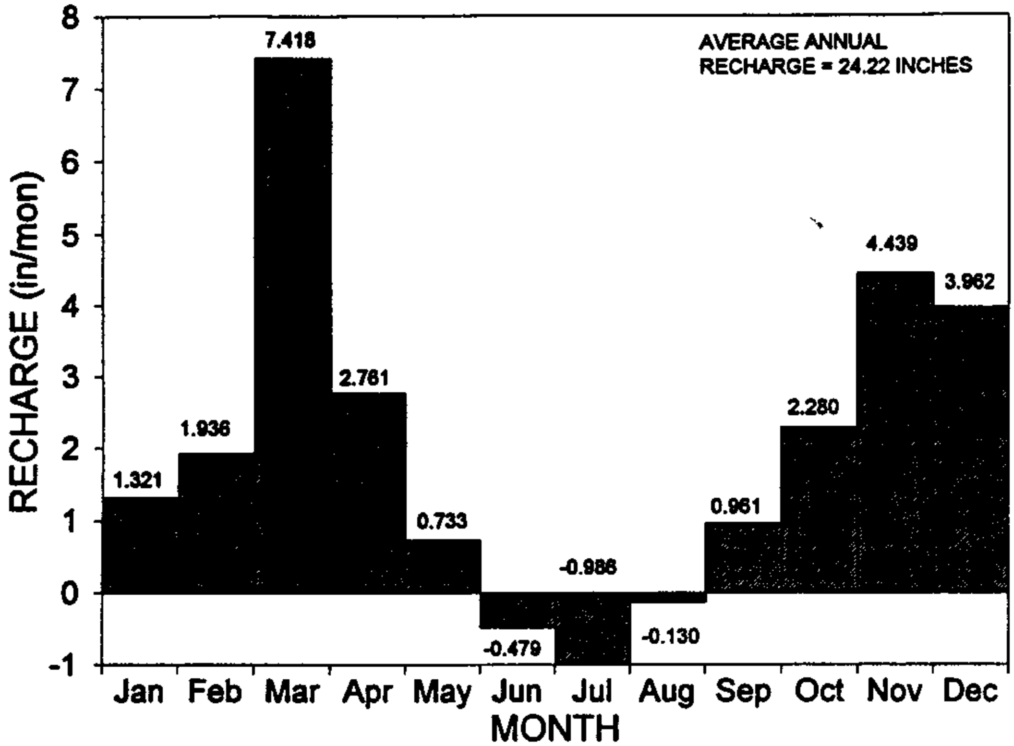

Figure 1. Average monthly recharge rates on sand and gravel deposits in Southern Maine.

tiometric head increase of 2.2 feet; multiplying calibrated recharge by a factor of 0.67 caused an average decrease of 1.5 feet in site area potentiometric heads. When all significant hydraulic parameters (such as till hydraulic conductivity) were varied one at a time by the same multiplication factors, the change in recharge rate caused the greatest change in site heads.

A final check on the recharge estimate was provided by measurements of vertical hydraulic conductivity. The Township 30 landfill site is on a local recharge area. With the large number of measurements of ground water vertical gradients on the site, it was possible to estimate the effective vertical hydraulic conductivity of the diamicton, which is approximately equal to the recharge rate divided by the average vertical gradient, or 0.04 feet per day.

RECHARGE RATE OF FINE-GRAINED LODGMENT TILL

If one knows the phreatic water table position and the hydraulic conductivities pertinent to a geologic cross section along a flow line with known boundary conditions at each end, the technique of specifying constant heads on the phreatic surface should provide the correct recharge into the section. Having a row of clustered multi-level piezometers on the section provides an important check on model calibration, particularly the hydraulic conductivity anisotropy which is one of the most sensitive parameters in cross section modeling. Large local fluxes can occur at sharp slope changes. In many field situations, a sharp change in slope is often associated with a discharge point that is at least seasonal in nature. Many of the model-simulated fluxes are created by juxtaposition of two constant head nodes on a steep slope, which causes a "short-circuit". Examining the flux below the top row of elements eliminates misinterpretation of the recharge rate that reaches deeper soil and rock zones.

Georgia-Pacific Landfill Study

An indirect method was used to estimate recharge to a thick silty glacial lodgment till at the Georgia-Pacific secure landfill site (RGGI, 1983). This site, a drumlin, is on the west bank of the St. Croix River and upstream of Woodland, Maine. A two-dimensional finite element model of the bedrock aquifer was applied to the regional flow system. A detailed finite-element cross section model was projected along an approximate flow line through the site. The models were calibrated with data from multi-level piezometers and insitu hydraulic conductivity measurements. The water table position in the cross section was known quite closely, and the stratigraphy and hydraulic conductivity of the till and bedrock units were well-documented. The phreatic surface in the model was fixed as a specified head boundary. With this boundary condition, the model calculated the flux into the till.

This was an important exercise as it also illustrated how recharge rates vary with proportional distance between the ground water divide and the discharge area. The averaged recharge rate over the soil cross section was 7.6% of precipitation (about 3.2 inches per year) but the localized rate was significantly higher (up to 25% of precipitation) on the top of the hill, then decreased in the downhill direction. These rates are taken as representative of soils with vertical hydraulic conductivities in the 0.015 to 0.0015 ft/day range with 35% to 60% passing a #200 sieve.

To extend the cross section results to the local plan-view two-dimensional model, the slope was divided into sections running parallel to ground surface contours, then an average recharge rate was applied in the slope sections within the same general slope position as calculated from the cross-section model. Of course, near the bottom of the slope, recharge rate drops to zero and discharge begins in streams and wetland areas. These discharge areas should be treated as a discharge point in the plan-view modeling, but not necessarily as constant head boundaries, since the streams and wetlands are not usually fully penetrating. Methods for simulating these 3rd-type Cauchy boundary conditions are given in most standard modeling references, e.g. Townley and Wilson (1980), McDonald and Harbaugh (1988), and Anderson and Woessner (1992).

Bald Mountain Hydrologic Study

A recent detailed hydrologic study for the Bald Mountain area of Aroostook County by (Fontaine, 1989a & 1989b) provides a baseflow analysis for watersheds comprised of shallow bedrock and thick, fine-grained glacial till. The till has 35% to 55% passing a #200 sieve and is derived from basic to intermediate volcanic rocks and graphitic shales. Insitu hydraulic conductivity values were typically in the range of 0.03 to 0.2 feet per day. Fontaine (1989a) calibrated a rainfall-runoff model (PRMS, Leavesley et al., 1983) to the Bald Mountain data by automatically varying recharge and other parameters. In addition to detailed meteorologic

Table 1: Annual Water Balance of Fine-grained Till in Northern Maine

Water Balance Component	Volume Equivalent	Percentage of Observed Precipitation
Observed Precipitation (P)	44.00 inches	100.0%
Interception (INT)	2.08 inches	4.7%
Net Precipitation (PNET)	41.92 inches	95.3%
Potential ET (PET)	21.28 inches	48.4%
Actual ET (ET)	13.81 inches	31.4%
Observed total runoff (ORO)	28.06 inches	63.8%
Water Balance Error (ERR1)	0.05 inches	0.1%
Predicted total runoff (PRO)	28.82 inches	65.5%
Surface runoff (SAS)	3.99 inches	9.1%
Subsurface flow (RAS)	17.75 inches	40.3%
Ground water flow (BAS)	7.08 inches	16.1%
Runoff mass balance err (ERR2)	-0.76 inches	-1.7%
Ground water reservoir inflow (RCH)	6.43 inches	14.6%
Ground water flow out of system (SNK)	-0.65 inches	-1.5%

Water Balance Relations:

P = ET + INT + ORO + ERR1
ERR1 = PNET - ET - ORO
PRO = SAS + RAS + BAS
RCH = BAS + SNK (SNK < 0: water into system)

PNET = P - INT
ERR2 = ORO - PRO

Reference: Table 15, Fontaine (1989a, p. 33)

and water quality measurements, detailed stream gaging was done on two forested watersheds with areas of 1.73 and 1.15 square miles, respectively. The PRMS calculates the various components of the catchment water balance as shown on Table 1. Neither Fontaine (1989a), nor this paper goes into detail on the fundamental topic of hillslope hydrology. However, one should consult a text such as Kirkby (1978) to gain an appreciation for the complexity of water movement from precipitation falling on the land to its final destination.

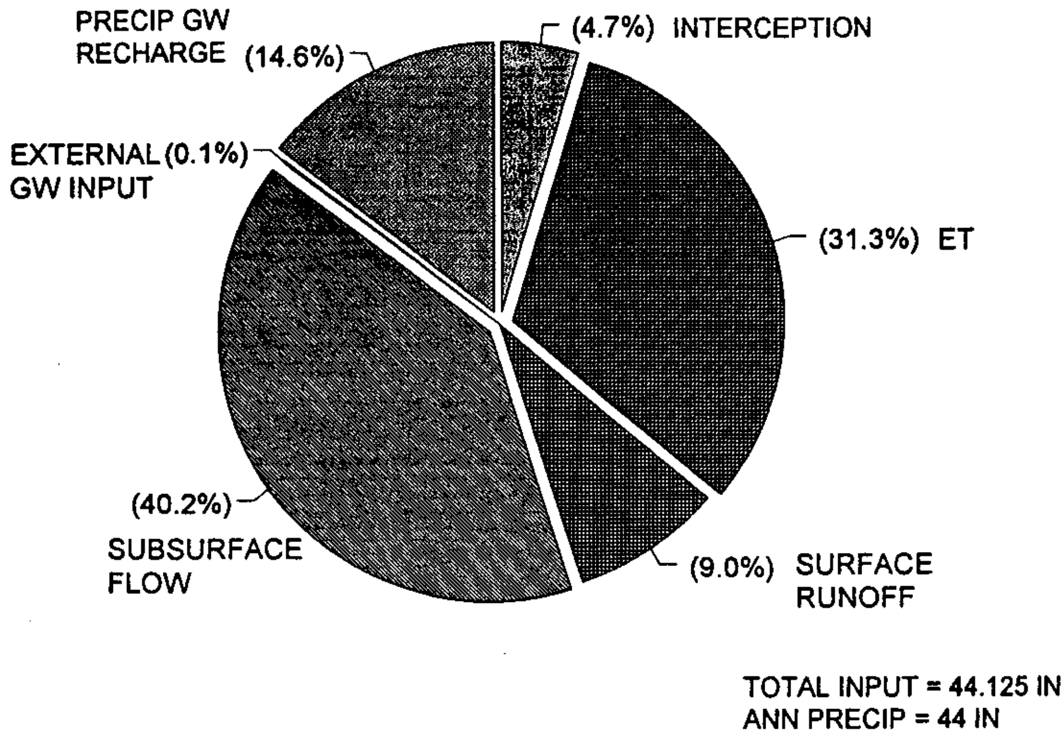

Figure 2. Annual water balance of fine-grained till in Northern Maine.

Fontaine (1989a) simulated the water year 1983 water balance with PRMS as given in Table 1. The same data is depicted graphically in Figure 2 (small discrepancies between percentages in Table 1 and Figure 2 are due to residual errors in the simulation). ET is a smaller percentage of precipitation than in the southern sections of Maine. The largest component of stream flow is the contribution from "subsurface flow" ("RAS" in Table 1), which is the shallow ground water flow taking place within a few feet of the ground surface. It is often a perched ground water table in the A- and B-horizons of the soil.

The ground water recharge rates discussed in this paper correspond to "ground water reservoir inflow", or "RCH", in Table 1. The deep ground water reservoir is below the zone of soil development and "subsurface flow". Twenty-seven percent of the Bald Mountain Brook watershed consists of steep slopes with thin soils, so the "ground water reservoir inflow" component, 6.43 inches per year, reflects a composite of both thick and thin till terrains. Simulated baseflow ("BAS"), or ground water discharge to surface water, is the sum of precipitation recharge to the ground water reservoir and any ground water exchanged directly with an external ground water system ("SNK"). Fontaine estimated a small (0.65 inches per year) net exogenous input to the ground water reservoir, for a baseflow of 7.08 inches.

Baseflow as estimated for Bald Mountain, 7.08 inches per year, is only slightly less than the estimated baseflow contribution for sandy till in Morrissey (1983) of 7.4 inches per year (19% of precipitation). These data suggest that there is only a minor difference in recharge potential among the tills at this site and those in the Little Androscoggin Valley and at the Township 30 landfill site. This finding is consistent with the similar till hydraulic conductivities at Bald Mountain and the Township 30 landfill site. Contrast this with the horizontal hydraulic conductivity of the Georgia-Pacific landfill site in Woodland, 0.0085 feet per day in the unweathered zone. Recall that in Woodland average recharge was estimated as 7.6% of annual precipitation, or 3.2 inches per year.

RECHARGE RATE TO GLACIOMARINE CLAY-SILT

The Presumpscot Formation clay-silt covers a large portion of the state of Maine (Thompson and Borns, 1985). It acts as an aquitard for bedrock and sand and gravel aquifers. It lies at the base of many of Maine's wetlands. It protects the ground water under the large commercial landfill in Norridgewock, Maine. The facility was formerly known as the Consolidated Waste Services (CWS) landfill, and is now called the WMDSM Crossroads Landfill. A knowledge of how much recharge moves downward into the soft "gray clay" zone is critical to contaminant dilution calculations and to ground water travel time computations. RGGI has developed recharge estimates for this ground water system by indirect calculation (1985) and manual and automatic inverse methods (1987b and 1993, respectively), i.e. calibrating by recharge to observed heads.

The Presumpscot Formation clay-silt is typically composed of about 10 feet of desiccated brown or olive clay-silt overlying a softer "blue" or gray clay-silt. The desiccated zone is fissured into a subangular blocky pattern, more dense and closely spaced at the ground surface and diminishing with depth. The softer gray clay lies below the position of the permanent water table. Because the fissured clay acts like a double porosity medium, ground water can rise rapidly in the fissures and is often at or near the ground surface in the spring. The hydraulic conductivity of the fissured zone is determined by the fissure pattern and can be about 50 times greater than in the unfissured gray clay.

The Crossroads landfill site has been studied extensively over the years by RGGI. Exhaustive hydrogeology studies and associated three-dimensional ground water model applications have been documented (RGGI, 1985, 1987, 1993). One of the most difficult parameters to estimate was the recharge rate of the clay-silt. Extensive data on the distribution and hydraulic conductivity of the clay-silt, underlying sandy till and bedrock was assembled. There were many clusters of multi-level piezometers throughout the site. Hydraulic conductivities were measured by numerous variable head tests in the piezometer clusters and several pumping tests in the bedrock.

The initial approach to modeling the site (RGGI, 1985) assumed that hydraulic conductivity and all other physical parameters of each of the 5 layers in the three-dimensional model were known. Recharge was treated as the only unknown and it was estimated indirectly by fixing constant heads in the top layer of the model, based on the known position of the phreatic surface. The model output gave the fluxes through the bottom of each model cell. This treatment is analogous to the fixing of the phreatic surface in the Georgia-Pacific cross section model described earlier, only at Crossroads Landfill (formerly CWS) the model was fully three dimensional. This approach indicated that an average recharge of 12.1 inches per year (29% of precipitation) entered the fissured clay, which was the top layer of the model. At Bald Mountain watershed in 1983 (Fontaine, 1989a), 24.1 inches of precipitation per year entered the silty till. Most of this recharge was presumably retained as shallow ground water flow.

The model was revisited in 1987 (RGGI, 1987). The hydraulic conductivities of the clay-silt, till, and rock were kriged according to hydrogeologic unit. The gridded data were used to develop spatially variable conductivity inputs for **MODFLOW**. Most of the constant head cells were removed from the model and instead direct recharge was applied to the model. The model was calibrated by trial-and-error, adjusting recharge until a good fit to observed water levels was obtained. Vertical hydraulic conductivity was also varied during this inverse procedure. After re-calibration of the three-dimensional model, it was determined that only 1.9 inches per year of precipitation (4.6%) moved down into the soft gray clay. This would be approximately equivalent to the mass flux vertically downward through a soil with a vertical hydraulic conductivity of 5 10-4 feet per day. This seems reasonable since the clay-silt hydraulic conductivities range from 3 10-5 to 3 10-3 feet per day, which is considerably lower than the values in the Bald Mountain till (horizontal Kh = 0.045 - 0.91 feet per day, vertical Kz = 0.5Kh, deep recharge 6.4 inches per year), and just slightly lower than the values in the till at the Georgia-Pacific landfill where deep precipitation recharge in the till (Kh = 8.5 10-3 feet per day, Kz = 1.4 10-3 feet per day) was 3.2 inches per year.

Further revision of the model was reported in 1993 (RGGI, 1993). The grid was expanded and recharge was estimated using the **MODINV** (Doherty, 1990) regression package for automatic estimation of **MODFLOW** parameters. The model was calibrated by fixing the hydraulic conductivity and varying the recharge. Considerable data on hydraulic conductivity were available: a total of 160 falling head tests, including 32 located in the bedrock, were available for estimation of hydraulic conductivity. Five bedrock pumping tests were also performed. Consequently, conductivity could be assumed known with some confidence. The computer program **MODINV** was used to perform the calibration by parameter optimization, which consisted of varying the recharge to minimize the RMSE head residual. Head observations from 201 individual piezometers comprised the calibration data set. There were two recharge zones: one representing areas with clay controlling the recharge, and the other representing areas with till controlling the recharge.

The calibrated recharge values derived using the optimization procedure were 0.53 inches per year (1.2 % of precipitation) for the clay zone, and 2.2 inches per year (5% of precipitation) for the till zone. These are total recharge values. Some of the recharge remains in the shallow flow zone and discharges locally through internal drainages, instead of passing through the gray clay to till and bedrock. The clay recharge is equivalent to the mass flux vertically downward through a soil with a vertical hydraulic conductivity of 1.2 10-4 feet per day (4.3 10-8 cm/sec). This agrees with the field hydraulic conductivity tests results, where the vertical hydraulic conductivity of the gray clay was also calculated to be 1.2 10-4 feet per day. These results are consistent with the 1987 results and are a refinement of that work.

A check on this work was provided by analysis using tritium age dating of water samples from the gray clay. To the authors' knowledge, this is the first application of age dating of ground water in Maine for the specific purpose of estimating recharge and vertical flux in overburden. Due to the lab method used (direct liquid scintillation counting), the data were not very precise (+/- 8 TU). The best results suggest a maximum recharge rate through the gray clay of 0.24 inches per year, assuming a porosity of 0.40. This value represents the portion of recharge that actually passes through the gray clay, and is therefore less than the total recharge value estimated by inverse modeling for the clay zone. This value is undoubtedly subject to some variation over the site, and also must be qualified by the fact that the precipitation tritium input function was developed using data from Ottawa, Canada. That said, the estimate is remarkably consistent with the results of automatic flow model calibration.

BEDROCK AQUIFER RECHARGE RATES

A more difficult problem is to estimate the rate of recharge to bedrock aquifers. Except for areas where bedrock is exposed at ground surface, all precipitation has to travel first through soil before penetrating into minute cracks in the bedrock surface. The relief of the land and the location of the point of interest with respect to ground water divides and discharge areas control the bedrock recharge rate. The thickness and effective vertical hydraulic

Table 2: Average Recharge Rates to Maine Bedrock

Project	Location	Rock Type/ Rock Name	Typical Transmissivity (ft2/day)	Recharge*
McKin Chemical	Gray	Sebago Pluton	100	7%
Sewage lagoons	Jackman	granite	40	4%
Aeration lagoons	Woodland	Cookson Formation (metasandstone) and granite	0.1 to 142	10%
Industrial landfill	Woodland	Cookson Formation	42	5%
Commercial landfill	Norridgewock	Sangerville Formation	150 NE to 50 NW	1-4%
Leaking gas tanks	Friendship	granite	15 ENE to 1.5 NNW	13%
CZM study	Portland Islands	metamorphic	43 NE to 4.3 NW	7.4-8.5%
Seawater intrusion	St. George	granite	0.6 to 200	10%
Seawater intrusion	York	Kittery Formation (quartzite)	75 to 400	5%

* expressed as a percentage of total average annual precipitation

conductivity of the soils are very important. Finally, the difference between the potentiometric surface in the bedrock versus the overlying soil aquifer controls the recharge/discharge rate. Local recharge rates can vary greatly depending upon all of these parameters. Nevertheless, some basic areally-averaged estimates that may be of use to hydrogeologists have been derived from detailed computer modeling studies performed by RGGI. These studies have involved three types of models:

- vertical cross section models along known or assumed flow paths in the bedrock aquifers;
- two-dimensional bedrock aquifer models that include the overlying soil as an "adjacent aquifer" with defined thicknesses, vertical hydraulic conductivities, and known water table positions;
- fully three-dimensional models.

In all of these studies model verification in the form of at least some bedrock water potentiometric surface data was available. In other cases pumping test data and/or insitu hydraulic conductivity tests were used to calibrate models by adjusting recharge. In all cases, field mapping of the density and orientation of bedrock fractures was performed.

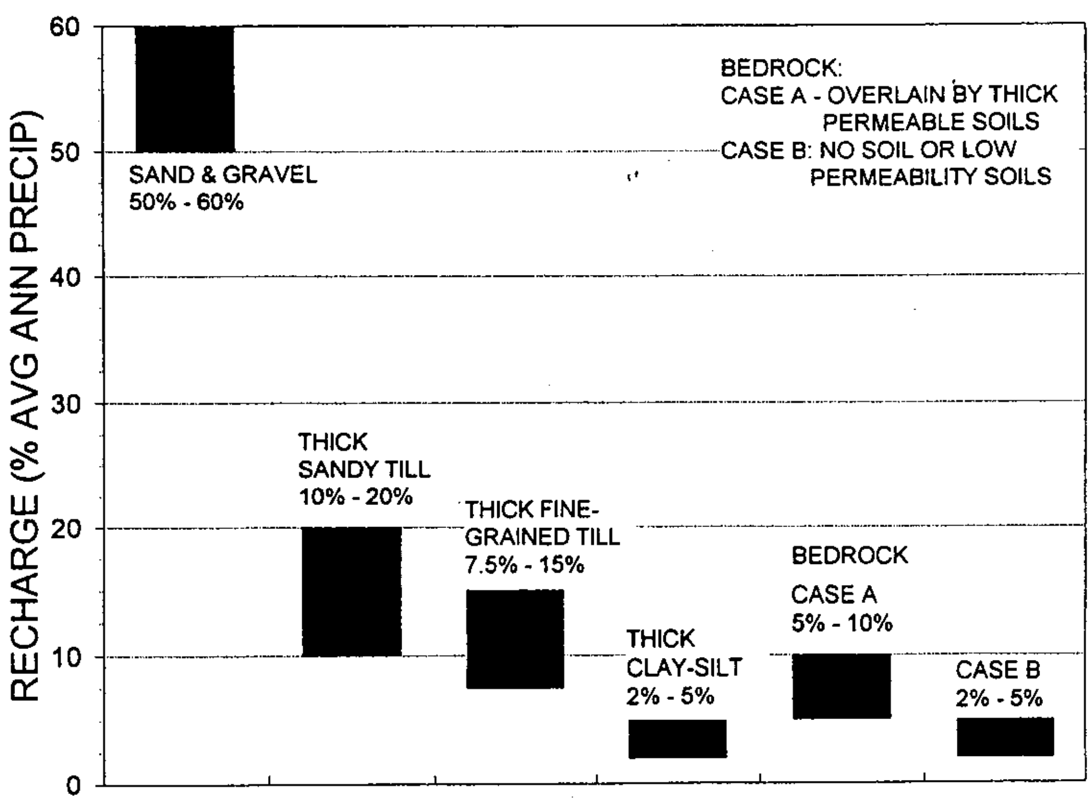

Figure 3. Suggested average ground water recharge rates according to geologic terrain.

Table 2 summarizes information from bedrock modeling studies by RGGI. In all cases, the results did not rely on a need to estimate the recharge rate to the soil, though other measured or estimated soil parameters were used (vertical hydraulic conductivity, soil thickness, and water table elevation in the soil). The sites listed on Table 2 with the lowest leakage factors (vertical hydraulic conductivity divided by thickness) in the overlying saturated soils generally have the lowest recharge rates. However, the site with one of the smallest recharge rates is the York site where there is almost no soil and therefore, no medium to hold water long enough for it to seep into cracks in the rock surface. In the York study (RGGI, 1988), a seawater intrusion model (Bear and Verruijt, 1987) was used which gave good agreement with historical saltwater intrusion in bedrock wells. The St. George model was also calibrated with many wells and much fracture data. The **AQUIFEM** model (Townley and Wilson, 1980) was used in that study, with the overlying soil treated as an adjacent leaky aquifer. The Friendship and Norridgewock models were calibrated partially through the use of pumping tests.

Although computer models are useful and indeed necessary to study bedrock recharge patterns in detail, the general recharge estimates given in Table 2 are probably representative of the range that exists in most Maine bedrock aquifers. Initial estimates for screening calculations and simple analytical models can be drawn from this compilation.

ADDITIONAL RESEARCH NEEDED

The modeling work by RGGI and the USGS has been time-consuming and costly, yet knowledge of effective recharge rates for many types of Maine geologic terrain is still incomplete. Additional detailed studies are needed on various surficial units to develop a broader recharge data base. One source of information might be other private-sector model applications that were unavailable to the authors. More detailed USGS water balance studies such as Fontaine (1989a,b) are needed on small watersheds of homogeneous surficial geology. To the authors' knowledge, the tritium dating at the Crossroads Landfill is the only age dating to have been performed in Maine for the specific purpose of estimating recharge. Age dating studies on different surficial units where downward gradients can be demonstrated are an important check on other estimation methods. More work is also needed on differentiating the change in recharge rate from ground water divide to discharge point. The seasonal variations in recharge for geologic units other than high water table sand and gravel aquifers need study. Finally, Maine-specific studies of hillslope hydrology are needed to analyze the division of water flow among the various types of flow within a slope. This would be particularly useful to the study of nitrate contamination from subsurface septic systems which are placed in the upper soil horizons.

SUMMARY

Ground water recharge rates cannot be measured directly through observation, and therefore it must be modeled in some fashion. Various methods of estimating ground water recharge have been reviewed. Examples have been presented of how these methods have been used in Maine to estimate ground water recharge at specific sites on different homogeneous geologic terrains. It is reasonable to believe that similar terrains within Maine have simi-

Ground water recharge rates cannot be measured directly through observation, and therefore it must be modeled in some fashion. Various methods of estimating ground water recharge have been reviewed. Examples have been presented of how these methods have been used in Maine to estimate ground water recharge at specific sites on different homogeneous geologic terrains. It is reasonable to believe that similar terrains within Maine have similar average annual recharge rates over large areas. Inverse and indirect modeling approaches are useful tools for estimating the distribution of recharge, but the models demand a good knowledge of hydraulic conductivity distribution, stratigraphic data, and bedrock fracture flow information. The detailed water balance studies coupled with models such as the USGS PRMS also offer promise, but are costly in terms of field data and data reduction requirements. Areas for investigation that would benefit all hydrogeologists have been suggested.

Studies with modest hydrogeologic budgets will not be able to estimate recharge independently. Therefore, recommendations for averaged annual recharge rates for the most common geologic terrains found in Maine are given in Figure 3. Some error will be involved in applying these uniformly over large regions, but the judicious use of 3rd-type boundary conditions can provide avenues of ground water discharge where appropriate. With some care, the monthly recharge rate distribution given in Figure 1 can also be adapted to other geologic terrains. Severe droughts can be approximated by applying 60% of the average annual recharge rates. In the absence of better data, enough information on the variables involved has been presented so that an informed hydrogeologist can adapt these estimates to most situations.

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