2.10 Changes in the lower boundary condition of water fluxes in the Noah land surface scheme

Dag Lohmann* and Ken Mitchell
Environmental Modeling Center
NOAA/NWS/NCEP
20746 Suitland, MD

1. Introduction.

The purpose of this work is to present a simple conceptual groundwater module that can be coupled to land surface models (LSM) used in weather prediction and climate models. The governing equations of this model are based on the linearized Boussinesq equations of groundwater theory. Subsequently the regional basin-scale effective parameters for this model are derived from streamflow measurements of a subset of 1145 river basins used in the North American Land Data Assimilation (NLDAS) project [Lohmann et al., 2003] with the help of the Brutsaert and Nieber [1977] recession slope baseflow analysis. Various authors applied this method regionally for different river basins [Vogel and Kroll, 1992; Brutsaert and Lopez, 1998; Troch et al., 1993] with contradictory results regarding the linearity of the baseflow recessions and therefore about the nature and complexity of long term basin behavior. This study aims to apply the techniques of these authors to basins within the NLDAS region which cover about 25% of the continental US. Also, the approach of Pauwels et al. [2002] will be expanded into a piecewise linear metahill-slope model and coupled to the Noah LSM. Model results for the retrospective period from Oct. 1996 to realtime will be shown. Here first results from the baseflow analsysis are presented.

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*Corresponding author adress:
Dag Lohmann
NOAA/NCEP/EMC Rm. 204
5200 Auth Rd., Suitland, MD 20746
email: Dag.Lohmann@noaa.gov

2. Description

Brutsaert [1994] derived the analytical solution for the linearized Boussinesq equation:

$$\frac{\partial h}{\partial t} = \frac{kpD\cos\Theta}{f}\frac{\partial^2 h}{\partial x^2} + \frac{k\sin\Theta}{f}\frac{\partial h}{\partial x} + \frac{R}{f}$$

under the assumption of spatially constant parameters, and the Dupuit-Forcheimer approximation (stream lines are horizontal and velocity uniform with depth); h is the elevation of of the groundwater table, k is the hydraulic conductivity, f the drainable porosity, p a parameter resulting from the linearization, D the soil depth, x the horizontal distance from the river bed, R the recharge rate, and Θ the slope of the aquifer. The solution for this equation can be written as the sum of simple exponentials in the case of a flat terrain where the second term on the right hand side vanishes. Kraijenhoff [1958] and Dooge [1973] notice that after an initial transient decay the solution looks like a simple linear storage equation whose parameters can be estimated from regression analysis. The parameter that determines the residence time of the linear storage was named reservoir coefficient.

To determine this coefficient and test the linear storage hypothesis, we performed a Brutsaert and Nieber [1977] baseflow analysis in 1145 basins within the US with areas from 25km^2 to 10,000km^2 from 1948 to 2002. An automatic hydrograph recession algorithm was employed to search for time intervals that define low flow periods. A recession is defined here as a period with at least 10 consecutive days of decreasing streamflow within the low flow periods from July to September. Only data after an initial transient of 7 days were analyzed. Figure 1 shows this alysis for one basin (see also Equation 12 in Brutsaert Nieber [1977]). The upper plot shows the measured streamflow (black) for the time period in 2002 that was used for the baseflow recession analysis (red). In the lower plot the x-axis is the measured streamflow, the y-axis is the negative time derivative of the streamflow for all recession periods. The slope of these to two quantities on a log-log plot determines the linear versus the nonlinear behavior of the basins.

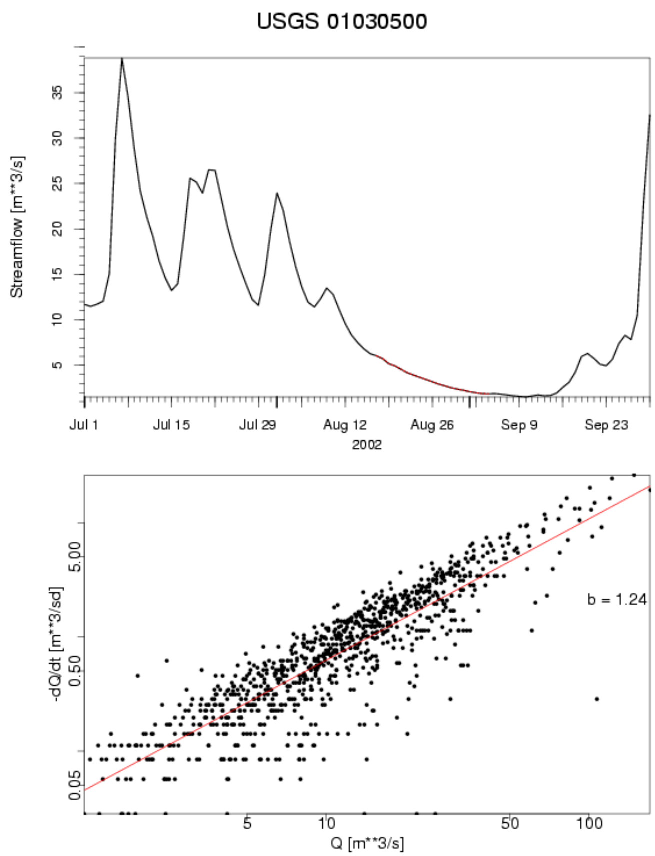

Fig. 1. Daily streamflow for USGS basin 01030500. The red line in the upper plot indicates the part of the measured streamflow that was used for the regression analysis. The lower plot shows the negative change of streamflow as a function of streamflow itself.The slope of the log-log line between streamflow and d(streamflow)/dt is 1.24.

Figure 2 shows a spatial plot of this regression coefficient. In general most of the slope values are around 1 to 1.5, which means that the linear assumption holds up in many areas. Based on these results we are encouraged to extract baseflow model parameters from measured streamflow that are consistent with the model that uses these parameters.

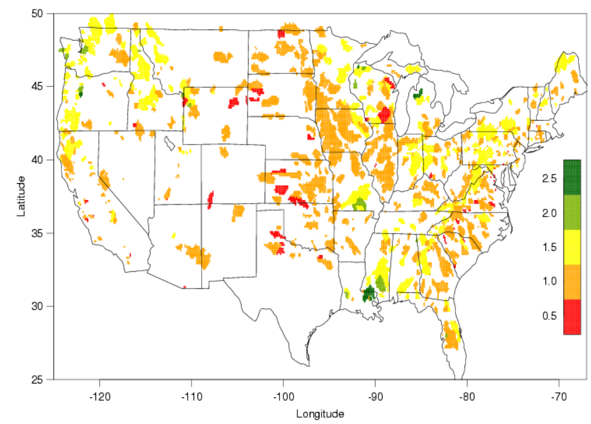

Fig. 2. Spatial distribution of the slope parameter b. Most slope parameter values are around 1 to 1.5.

3. Model development

Pauwels et al. [2002] showed that a simple model based on the linearized Boussinesq was able to reproduce baseflow in a small catchment in Belgium. They build on the conceptual model of Brutsaert [1994] and advanced it by including a variable recharge rate. A similar model will be extended into a piecewise linear metahillslope model that can be adapted to the current grid setup which is common to atmospheric models and LSMs [e.g. in the NLDAS project, Mitchell et al., 2003]. Each grid cell will be sub-divided into areas of similar baseflow responses. A linear routing model will be coupled to these modules based on linearizations of the St. Venant equations [Lohmann et al., 2003]. Also, alternative previous routes will be investigated like Topmodel based approaches and full 3-d groundwater models. These will be analyzed with the baseflow recession techniques used in this paper.

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