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Hydrologic Trends in the Middle East: Modeling and Remote Sensing

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Abstract: In this report, we use several tools to document recent large scale trends in Middle East hydrology. Four interelated aspects are investigated: 1) upland snow cover in the Zagros and Taurus mountains, 2) runoff and discharge in the Tigris, Euphrates and other rivers, 3) rainfed agriculture in the Fertile Crescent and 4) irrigated agriculture using river waters. The primary tools used in this work are conventional climate data sets, a 1982 to 1998 time series of composited AVHRR satellite data and a simple hydrology model run with a monthly time step on a 5 kilometer grid. We show that interannual variation in winter temperature has a significant influence on the extent of winter snow cover by shifting the snow line. The hydrology model, including snowmelt, predicts the timing and magnitude of river discharges. Watersheds at different elevations discharge a fraction of rainfall ranging from 10 to 60 percent.

Statistical correlation of climate and satellite data indicates that winter minimum temperature influences the density of April and May rainfed crop vegetation in the northern border region of the Fertile Crescent. Winter precipitation controls the density of the April, May and June crop density in the Fertile Crescent and in marginal lands on the steppe. Spring rains control natural steppe grasses in April and May.

Multi-scale analysis of AVHRR and Landsat images allowed us to estimate the trend in irrigated area. Rapid increases in irrigated lands occurred between 1992and 1998 in large new “plantations”, in traditional river bank locations and in thinly distributed areas in Mesopotamia. Recent growth in irrigated land has increased evaporative water losses to levels comparable with river discharges. Along the lower Balikh, irrigation has decreased due to upstream withdrawals.

In the final section, the model is used to predict the sensitivity of river discharge and rainfed agriculture to cooler/warmer, wetter/dryer conditions.

1. Introduction

The climate and hydrology of the Near East, including the Fertile Crescent, are of both theoretical and practical interest. From a theoretical point of view, the region poses special challenges to the climatologist due to the impressive Tauros and Zagros mountains along its northern rim, its diverse water bodies (Mediterranean, Black and Caspian Seas and the Persian Gulf) and the strong north-south climate gradient. Practically, the climate of the region is of interest due to its historical role in the development of agriculture, irrigation and civilization (McCorriston and Hole, 1991; Nissenbaum, 1994, Hole, 1994) and modern concerns about rapid population rise, water resources and political conflict. [Some general data about population and agriculture is given in Table 1.]

A general description of Near East climate is given by Taha et al. (1981). More detailed country-specific climate atlases are available in national reports. The general relationship between climate and agriculture is presented by de Brichambaut and Wallen (1963) and Smith and Harris (1981). Comprehensive analyses of natural vegetation and climate are given by Nahal (1981), Zohary (1973), Bakours and Kolars (1993) and in FAO reports ( ).

The impact of climate change on Early Bronze Age civilizations in the Near East (2300 to 1900 BC) is discussed by Weiss (1993), Gibbons (1993) and Courty and Weiss (1997). Impacts on the Late Bronze Age is analyzed by Weiss (1982), including some reference to modern patterns of interannual variability. Proxy data for Tigris-Euphrates discharge over the second half of the Holocene (i.e. 6000 ybp to today) is analyzed by Kay and Johnson (1981). Climate simulations for this period are described by Kutzbach et al. (1996). Recent century-scale trends in climate have been examined by Cohen and Stanhill (1996) and on a larger scale (but with local implications) by Ropelewski and Halpert (1987). The issues of Near East water resources, population growth and political conflict are discussed by Al-Kashab (1958), Beaumont (1996, 1997a,b, 1998), Kolars (1992, 1994), Tvedt (1992), Bilen (1994) and Wolf (1994).

The objective of this report is to develop quantitative estimates of trends in aspects of Middle East hydrology. We include a large area in our study, in excess of a million square kilometers, including eastern Turkey, Syria, Lebanon, Iraq, northern Iran, Israel, Jordan and northern Saudi Arabia. These countries share large watersheds and face common challenges in regard to rainfed and irrigated agriculture. Our analysis also extends to small spatial scales. We must resolve tight climate gradients associated with coastlines and mountain slopes. In estimating trends in irrigated agriculture, fields no larger than a 100 meters must be resolved.

The paper is divided into two parts. The first half, including sections 2, 3 and 4, describes the study region and our methodology. Section 2 describes the terrain and climate of the study region. Section 3 derives a simple hydrology model applied to a 5 *km* grid. Section 4 provides a survey of satellite derived land use and land cover and gives a glimpse of the extent of recent landscape change.

The second half of the paper (Sections 5, 6, 7 and 8) analyzes four interrelated aspects of Near East hydrology: mountain snow cover, river discharge, rainfed agriculture and irrigated agriculture. These results are summarized in section 9 and in section 10, we consider certain climate change and anthropogenic change scenarios using the sensitivity coefficients derived in the earlier sections.

[Table 1]

2. Geography and Climate Distribution in South West Asia

The region included in this study extends from 32°E to 51°E in longitude and 30°N to 41°N latitude. A large portion of this area is shown in Figure 1, along with terrain height, country boundaries, watersheds and FAO/WMO climate station locations. This subregion is of importance for our study as it includes the Fertile Crescent and several historical and modern centers of population and agriculture. An additional northern section (not shown in Figure 1) is retained in our study as it contains the high mountain watersheds that supply the Euphrates and Tigris Rivers and underground aquifers that supply irrigation water to the region.

The monthly temperature and precipitation climatology of the Near East are shown in Figures 2, 3, 4 and 5, derived from the FAO/WMO stations for the period 1955 to 1972???. This data is interpolated to a five kilometer grid as described in Appendix 2. The interpolation scheme includes a $-5°C/km$ temperature lapse rate correction but no corresponding correction to precipitation. A review of previous work (Alpert and Shafir, 1989, 1991; Basist and Bell, 1994; Gobel et al. 1996; Hevesi et al., 1992; Frei and Sch[illegible]r, 1998; and others) indicated that simple altitude and aspect-based precipitation corrections are not sufficiently accurate to add information and confidence to our fields. Our model also neglects the statistics of daily rainfall (Dennet et al., 1984).

For conciseness of presentation, the seasonal and spatial distributions of climate have been classified. In Figures 2 and 3, each class contains pixels with similar average values and season cycles. The mean seasonal cycle for each class is plotted in Figure 4. These class maps and signature plots capture most of the climate features found in previous manual and computer analyses of Middle East climate (de Brichambaut and Wallen, 1963, Taha et al., 1981). As seen in Figure 2 and 5, the temperature increases from the higher elevations in the Tauros and Zagros mountain areas in the north to the deserts of southeastern Iraq. The annual range of temperature is about $25°C$ throughout the domain (classes 1-8) except for the maritime influence in Class 9 along the Mediterranean Sea with a range of about $15°C$. The finer spatial details of the temperature field seen in Figure 2 are not resolved by the existing station distribution but arise instead from the lapse rate correction. Temperature classes 1-4 have extended periods of sub-freezing temperatures in winter.

The spatial and seasonal distribution of precipitation (Fig 3, 5) shows a strong increase of precipitation northward and outward from the deserts of Saudi Arabia, eastern Jordan, western Iraq and southeastern Syria toward the mountains and the Mediterranean and Caspian Seas. The sequence of precipitation classes encountered along a south-to-north S transect at 40°E (Classes 1, 2, 3, 5, 6) have annual precipitations amounts of 120, 200, 330, 500 and 700 mm. Somewhat less precipitation (Class 4) is found in the higher mountains with local maximum seen along Iraq's northern border with Turkey and Iran (Class 8). Classes 7 and 9 , along the Levantine Coast, have a sharper January maximum than the other classes. Most of the region experiences a winter maximum of precipitation characteristic of Mediterranean climates (Fig. 3a). One exception (Class 10) has a fall-season maximum along the Caspian Sea.

3. A Hydrology model

To convert the average monthly FAO climate data to more useful information on soil moisture, snow cover, growing season and river discharge, a simple bucket hydrology model was used (Thornthwaite, 1948). This model, run on a 5 *km* grid, considers precipitation, actual and potential evapotranspiration (*AET* and *PET*), soil moisture storage, runoff, and snow. It uses a temperature dependent *PET* and assumes that the *AET* is proportional to the soil moisture and the *PET*. Coefficients of the model are given in Table 2. Typical inputs and outputs of the model are shown in Table 3. The model is driven by the spatially interpolated average monthly values of temperature and precipitation (Figs 2,3,4). The mathematical properties of the model are given in Appendix 2.

[Tables 2 and 3]

The computed seasonal cycle of soil moisture is shown in Figures 3 and 5. In most of the region (classes 1-6, 7-9) soil moisture reaches a maximum in February, about a month later than the time of maximum precipitation. In Class 10 along the Caspian Sea, soil moisture rises more quickly in the Fall due to the Fall pulse of precipitation mentioned above (i.e. precipitation class 10). Another interesting anomaly is Class 7 in the high valleys of the northern Iraq. In these dry cold regions, the Fall season cooling drops the temperature to below freezing (Fig 2, class 3,4) before the increasing precipitation (Fig 4, class 4) can saturate the soil. Thus, the model predicts a static dry frozen soil from November through February while snow accumulates. The melting snow in March immediately saturates the soil which then slowly dries again during the Spring.

Two annual quantities are useful in diagnosing the model and understanding the variability of climate in the Near East: “Wetdays” and Annual Runoff. The quantity “Wetdays” is defined as the number of days for which the soil moisture exceeds a critical value (e.g. SM_{crit}=50 *mm*). The wet period begins in the late fall when *SM* first exceeds SM_{crit}, and ends in the late spring when *SM* drops below SM_{crit} (see Smith and Harris, 1981). For example, in Figure 4 we see that class #2 has no soil “wetdays” (WD=0) while class #3 has WD=75. The spatial distribution of Wetdays gives an approximate idea of where soil moisture exists over a sufficient period of time to support the growth of natural vegetation and rainfed crops. More complete models of climate-biosphere interactions are described by Foley et al. (1996).

A sub-sampled scatter plot showing the relationship between Wetdays and annual precipitation is shown in Figure 7. The considerable scatter in this figure indicates that annual precipitation alone is not a good predictor of Wetdays. For example, for an annual precipitation of 400 *mm*, the value of wetdays varies from 100 to 2500 days. The abrupt rise in Wetdays from zero to 100 days as precipitation rises from 150 to 200 *mm* agrees with the conventional wisdom of a threshold value of 200 *mm* for rainfed griculture in Mediterranean climates.

A well known moisture index proposed by Koppen and Geiger (Oliver 1973) is the ratio of annual precipitation (*cm*) to annual average temperature (°*C*). We define the inverse tangent of this ratio as the Koppen-Geiger Mediterranean Index (*KGMI*), expressed in radians. That is

$$KGMI = \arctan(P(cm)/T(\circ C)) \qquad (1)$$

Koppen and Geiger suggest that values of *KGMI* > 1.11 give wet climates; values between 0.785 and 1.11 give steppe and values less than 0.785 give desert. Figure 6b shows that the use of *KGMI* collapses the scatter in predicted WD shown in Fig 6a. This result is not surprising as our formulation for potential evaporation (Appendix 2) guarantees that the ratio *P*/*T* is nearly proportional to the ratio of annual precipitation to annual potential evaporation. The arctangent transformation in (1) is also convenient as it gives a nearly linear relationship between KGMI and Wetdays, beyond the threshold at KGMI = 0.7 radians (fig. 6b). That is

$$WD = 300 \times (KGMI - 0.7) \qquad (2)$$

Without the arctangent transformation, the relationship between P/T and Wetdays is highly curved and more difficult to characterize. From Figure 6b and (2), the critical value of *KGMI* = 0.785 corresponds closely to the zero threshold for Wetdays and the value *KGMI*= 1.11 corresponds to Wetdays = 120 days, a reasonable value for the moist boundary of the steppe. This agreement with earlier work adds confidence to our hydrologic model and our choice of SM_{crit}.

Another important annual quantity is the total runoff for each surface element. Runoff is significant for the computation of river discharge (see Section 6) and for estimates of fluvial erosion. As seen in Figure 3, Soil Moisture classes 1-4 never saturate and therefore generate no runoff. Soil Moisture class 5 saturates briefly in January and February while classes 6-9 exhibit a longer period of saturation leading to significant runoff in the winter months. A scatter plot of runoff versus annual precipitation is shown in Figure 6c. The scatter is considerable, due to the variable influence of temperature. In general, no runoff is generated at locations receiving less than 300 *mm* although a 450 *mm* threshold is seen in some warmer locations. The slope of the scattergram is close to unity, indicating that for regions exceeding the threshold, every extra amount of annual precipitation give an equivalent extra runoff. The use of *KGMI* does not collapse the scatter shown in Figure 6c.

4. Land Cover

An important step toward our goal of understanding Middle East landscape and hydrology is the classification of land cover and land use (*LC*/*LU*). Unfortunately, this process is inherently difficult and uncertain. For the present purposes, we have used a method of multi-temporal NDVI classification, used recently by the USGS in their global *LC*/*LU* map (Defries and Townshend, 1994; Moody and Strahler, 1994; Gutman et al., 1995). This method groups together pixels which have a similar seasonal cycle of vegetative cover into “classes”. The vegetative cover in each 10-day period is measured by Normalized Difference Vegetation Ratio (NDVI), defined using the ratio of reflected near infra-red to visible radiation. The results, computed on a 1 *km* grid for 1993, are shown in Figure 7. In this figure, the Near East region is divided into thirteen vegetation classes using an unsupervised *k*-means ISOCLASS clustering algorithm operating on a February-to-September set of AVHRR NDVI images (Kouchoukos, 1998). The decision to identify just thirteen classes was made as a compromise between excessive lumping and splitting of classes. For convenience, we refer to this particular classification as SWAP13/93. The class-mean seasonal cycles, or “class signatures” for SWAP13/93 are shown in Figure 8. Legends for the thirteen classes are given in Table 2.

Classes 1 and 2 are desert regions with class 2 having a slight greening in the spring. Classes 3, 12 and 11, lie in northward-progressing concentric bands comprising the Fertile Crescent. These Classes exhibit a rapid rise in vegetation in the spring, peaking sharply in April, May or June respectively. Class 3 vegetation drops off in the summer while 12 and 11 maintain some vegetation throughout the dry period. This signature is characteristic of winter grain farming in Mediterranean climates.

Classes 5, 8 and 9, located in higher terrain and colder climates, also show a rapid vegetation rise in spring, peaking sharply in June, dropping by half in July through October, then dropping further in November indicating barren land or snow. Class 7 is mostly temperate deciduous forest, with leaf area rising rapidly in spring, peaking in June/July but remaining high through October. Classes 4 and 6 lie at high altitude. Their vegetation is similar to Class 7 but with less vegetation, and drops to very low values from Dec to April, indicating snow cover (see Section 5). Finally, Classes 10 and 13 (possibly part of Class 9 as well) are mostly irrigated agriculture with crop leaf area peaking in both April and September (see Section 7).

While the satellite derived classification SWAP13/93 has not yet been extensively ground truthed, several aspects of its accuracy can be discussed. First, we have compared SWAP13/93 with the global 1993 1 *km* land cover classification produced at the USGS by Loveland et al. (19??). In the Near East region, the USGS product appears to be inferior to SWAP13/93. It has considerable spatial noise in the form of hundreds of isolated pixels which don’t match the surrounding areas. Furthermore, many of the class legends are inappropriate for Middle East land cover types, a result of its global character.

Both SWAP13/93 and USGS tend to misidentify regions of irrigated agriculture, in the following sense. The same classes that nicely capture irrigated agriculture (9,10,13) also include some natural landscapes further north. The reason for this confusion can be traced to the weight given by ISOCLASS to the mean and variation of seasonal NDVI signatures.

The predicted length of the moist season (i.e. WD) is overlain on the SWAP13/93 as contours in Fig 7. As expected from previous work, there is considerable correspondence between the *WD* contours and the SWAP13 class boundaries. The desert classes (1,2) have *WD*< 125 days, The dry steppe with marginal barley cultivation (Class 3), generally lies between *WD*=125 and 150 days. The northern edge of the steppe (Class 12) lies roughly between *WD*=150 and 175 days. In several locations, the *WD* contours run almost precisely parallel to the class boundaries. Noting that the network of climate stations is insufficiently dense to allow this type of correspondence, we conclude that the temperature lapse rate correction, acting through the evaporation formulation in the hydrologic model will often force the *WD* contours to follow topographic contours. If the actual vegetation also respects the topographic contours, the *WD* and class boundaries will occasionally be parallel.

The most obvious discrepancy between *WD* and vegetation are the irrigated areas (i.e. Classes 9, 10 and 13) in which vegetation does not depend on local rainfall. Most striking in this respect are locations along the Euphrates, the Tigris and in Mesopotamia.

A multi-temporal NDVI-based land cover classification has also been carried out for 1995 (i.e. SWAP13/95, not shown). The 1993 and 1995 class maps and NDVI seasonal signatures are qualitatively similar but show some significant differences in detail. The statistical similarity between SWAP13/ 63 and 95 was computed using standard methods (Congalton, 1991). After matching classes, the number of pixels falling in the same class in both years is 39%. To evaluate the effect of the number of classes, vegetation patterns in both years were reclassified using *N*= 8 and 5 instead of *N*=13. The number of common pixels were then 50% and 62% respectively; still relatively poor agreement. While this lack of correspondence between 1993 and 1995 could be due in part to satellite remote sensing errors coupled to the mathematical sensitivity of the ISOCLASS algorithm, we believe that most of the discrepancy is due to actual land cover changes. As we shall show, land cover in almost every pixel in the Middle East is subject to interannual variation in at least one of the following conditions: snow cover, deciduous tree phenology, crop frost damage, annual grasses, cropland rainfall, river water for irrigation and economic/policy impact on agricultural practice.

5. Mountain Snow Cover

The significant role that mountain snow plays in water storage and release makes it a relevant basis for detecting the influence of interannual climate fluctuation and for testing our hydrologic model. The region selected for snow analysis is a large rectangle with upper left coordinates (42:49N, 31:51E) and lower right coordinates (32:01N, 49:00E). This region includes the upland sections of the Euphrates and Tigris watersheds as well as those of other rivers draining eastern Turkey and northern Iran. Using the 13-year 8 *km* AVHRR composites, the monthly snow cover was determined based on combination of reflectance of visible radiation (channel #1) and the emission of thermal infrared radiation (channel #2). Similar techniques have been used on a broader geographical scale by Walland and Simmons (1997). On a smaller scale, Gil'ad and Bonne (1990) used Landsat data to estimate the contribution of snow to Jordan River runoff on Mt. Hermon (35:50E, 33:25N).

The spatial pattern of snow cover for the month of January is shown in Figure 9. As expected, this distribution exhibits small scale patterns that reflect the higher terrain elements seen in Fig 1. The large areas of intermediate February snow frequency indicate the that snow cover fluctuates considerable from year to year. A better view of this impressive interannual fluctuation is given in Figure 10, where the winter-season snow cover history is plotted independently for each year from 1983 until 1994. In 1985 the maximum snow cover was only

$180 \times 10^3\ km^2$ while in 1993 the snow cover reached $570 \times 10^3\ km^2$. The 12-year-average monthly snow cover has its seasonal maximum in February with a value $330 \times 10^3\ km^2$.

Figure 10 provides an opportunity to evaluate the snow predictions of the hydrology model, keeping in mind that the model reference run uses only an average monthly climatology from an earlier period (i.e. 1955-1975). The predicted total snow cover area agrees with the satellite observations reasonably well, but the model underpredicts in the early winter and overpredicts in the spring.

A more detailed comparison of predicted and observed snow cover is shown in Figure 11. In this figure, a four color scheme is used to compare the area of model predicted snow with the area observed to have snow in at least 6 of the 12 observed years. Black shading indicates that snow was neither predicted nor observed. The area in yellow was both predicted and observed to have snow. Green indicates that snow is observed but not predicted. Snow was predicted but not observed in the red area. Some systematic errors are evident. The model slightly overpredicts snow around the perimeter of the high mountain areas while significantly underpredicting snow on the mountain slopes facing the Caspian Sea. This latter discrepancy could indicate that during precipitation events in this area, northeasterly winds from Central Asia have lower air temperatures than the monthly averaged values used in the model calculation. Thus, snow would fall when rain is predicted.

In order to determine the sensitivity of snow cover to changes in winter temperature, we define two indices: the average snow cover (*A*) and the temperature (*T*) for December-January-February. The temperature index is computed from three clusters of six highland climate stations in western, central and eastern Turkey. The centroids of these station clusters lie along a latitude of 39.05N at longitudes of 31.9, 37.5 and 41.3 E. These indices are plotted against each other in Figure 13. The correlation of snow cover with *T*2 and T3 is rather good (i.e. R^2 values of 0.52 and 0.63), even though these station clusters lie to the west of the central snow covered area. The correlation between snow cover and *T*4 is poorer. The slope of the T2 and T3 regression lines are

$$\frac{dA}{dT} = -48\ and\ -49 \times 10^3\ km^2\ C^{-1} \qquad (1)$$

A similar snow/climate sensitivity calculation can be done using the hydrology model. We altered the reference run by increasing and then decreasing all input temperatures by 5°C. The modified winter growth and decay of the snow cover are shown in Figure 12. The predicted DJF-averaged snow cover are $33 \times 10^3\ km^2$ and $537 \times 10^3\ km^2$ for the warmer and colder runs respectively, compared to $170 \times 10^3\ km^2$ for the reference run. The snow cover sensitivity is then

$$\frac{dA}{dT} = \frac{33 - 537}{10} = -50 \times 10^3\ km^2\ C^{-1} \qquad (2)$$

The agreement between observations and model is remarkably good, perhaps because of the inherent simplicity of snow cover dynamics in steep complex terrain. It suggests that with a 5 *km* grid cell, we have properly resolved the terrain elements that contribute to snow cover and that the lapse rate of $\gamma = -5°C/km$, derived from sparse station data, is approximately correct.

In order to explain the value of snow sensitivity found above, we hypothesize that the snow area is controlled by "snow line dynamics" whereby interannual variation in DJF temperature simply alters the altitude of the snow line. Two parameters enter this theory: the atmospheric lapse rate (γ) and the cumulative terrain histogram ($A(z)$). The quantity $A(z)$ is defined as the land surface area above the altitude z. If each degree of warming (dT) lifts the snow line $dz = \gamma^{-1} \times dT$;

$$\frac{dA}{dT} = \gamma^{-1} \times \frac{dA}{dz} \tag{3}$$

where dA/dz is the non-cumulative terrain histogram. As shown in Figure 13, dA/dz decreases with altitude. For the range $z = 1$ to 4 *km*, we fit the cumulative histogram with

$$A = a \times Z^2 \tag{4}$$

where Z (km) is the distance measured down from the approximate terain peak at 4 *km* (i.e. Z=4 – z) and the coefficient $a = 97\times10^3$ (no units). The non-cumulative histogram is then (using 4)

$$\frac{dA}{dz} = 2 \times a \times Z \tag{5}$$

This representation of the terrain histogram is “conical” in the sense that if all the terrain elements where collected into a single axisymmetric peak, its shape would be a cone with radius $r=(a/\pi)^{1/2}\times Z$. The surface slope of this cone is $(\pi / a)^{1/2}$ = 0.0057.

If the mean snow area in DJF is $A = 270\times10^3\ km^2$, (4) gives the snow line altitude of Z=1.67 *km* and z=2.33*km*. From (5), $dA/dz= 324\times10^3\ km$. Using a lapse rate of $\gamma = -5°C/km$ (3) gives

$$\frac{dA}{dT} = \frac{324 \times 10^3\, km}{5\, C\ km^{-1}} = -65 \times 10^3\, km^2 C^{-1} \tag{6}$$

This sensitivity value is somewhat higher than the observed value (1) and the model prediction (2). Using the model-predicted mean DJF snow area of $170\times10^3\ km^2$ gives, using (5,6), Z= 1.32 *km* and

$$\frac{dA}{dT} = \frac{257 \times 10^3}{5\, C\ km^{-1}} = -51 \times 10^{-3}\, km^2\ C^{-1} \tag{7}$$

The sloping histogram (Figure 14) explains the marked asymmetry in snow cover response between the 5 degree warming and 5 degree cooling of model temperatures (see Fig 10). The 5°C warming gives a smaller sensitivity of; dA/dT= (33-170)/5 = $-27\times10^3\ km^2$/°C , because the last remnants of upland snow are nearly removed. Cooling gives a larger sensitivity; dA/dT =(537-170)/(-5) = $-73 \times 10^3\ km^2/°C$, as the extensive midlands begin to become snow covered.

6. River discharge

The sensitivity of river discharge to climate change is of obvious significance to to the Middle East region, but remote sensing provides no direct method for its determination. We proceed to use the hydrology model described in Section 2 and Appendix II, emboldened by its ability to predict the boundaries of the steppe (Section 4) and the amount and climate sensitivity of snow cover (section 5).

As a first step, the USGS Gtopo30 Digital Elevation Model (DEM) was used to determine the watershed boundaries for each of the major rivers in the region (Fig 1). For each watershed, the monthly model-predicted runoff values for each 5-by-5 *km* cell were added and assigned as the discharge value. No time delays associated with surface, soil or streamflow were considered. As in the snow study, the hydrology model was driven by average monthly climate data from the period 1955 to 1977. The results for the Euphrates, Upper Tigris and Greater Zab are shown in Fig 14. Annual total discharges for all the computed watersheds are given in Table 5. Figure 14 and Table 5 also include observed river discharges from earlier periods available in the literature. More modern discharge data are difficult to obtain and less appropriate in the present study, due to the extensive dams and diversions existing along these rivers today (e.g. Bilen, 1994).

Consider first the reference run for the Euphrates River shown in Figure 15a. The discharge reaches its maximum values of 2800 and 3000 m^3/s in March and April due to the rapid melting of snow (see Figure 10). The discharge data from the gauge at Hit shows maximum discharge values which are equal to or smaller than model values, occurring in April and May. The model and observations also differ in the summer months. The model discharge drops to zero while the actual discharge levels off near 300 m^3/s. Model errors in time of maximum flow and the level of summer flow are presumably both due to the lack of a water storage and delay mechanism in the model. The annual total discharge from the model (27048 *mcm*) and the two Hit records (22398 and 38438 *mcm*) are in rough agreement.

The Greater Zab comparison gives similar results. The model is too early in the time of maximum flow and too low in summer discharge, but it gives good agreement on annual discharge. The Upper Tigris comparison is a little worse. The model-predicted annual discharge is slightly lower than the Mosul observation and the model barely captures the spring discharge peak. This is probably because the model is using winter temperatures in the Upper Tigris watershed which are too warm to hold the water in the frozen state over the months of January and February. As seen in Table 5, model predicted annual discharges for the Little Zab and the Ceyhan are about 20% lower than the available observed discharges.

[Table 5]

The model-derived climate sensitivity of the Euphrates, Upper Tigris and Greater Zab river discharges are shown in Figure 14. Consider first the Euphrates River. An increase or decrease in precipitation by 25% raises or lowers the discharge profile while keeping its shape unchanged. The annual discharge rises to 40655 *mcm* or drops to 15751 *mcm* compared to the reference value of 27048 *mcm*. This is a 50% rise and a 42% drop, nearly twice the imposed percentage change in precipitation. The factor of two in relative sensitivity is due to the fact that runoff, which feeds discharge, is a threshold phenomenon. Below a certain amount, precipitation is balanced by evaporation. Once the threshold is exceeded, a proportional increase in precipitation causes a super-proportional increase in discharge.

An imposed change in temperature changes both the shape and magnitude of the Euphrates discharge. A 5 degree warming increases evapotranspiration thus lowering the discharge curve dramatically, dropping the annual discharge from 27048 *mcm* to 16329 *mcm*. The warming also eliminates the spring peak by preventing the overwinter storage of water in the mountain snow pack. A 5 degree cooling increases the annual discharge to 38607 *mcm*; amplifying and delaying the spring peak by almost a month. Similar sensitivities to temperature change are seen in the Upper Tigris, Greater Zab (Figure 15 b, c) and the other rivers (Table 4).

7. Irrigated Agriculture

7.1 Spatial distribution

It is generally known that there have been significant changes in the amount and distribution of irrigated agriculture in the Middle East in recent years, driven by population increases and changes in agricultural policy and economic factors (Table 1). However, considerable uncertainty exists in the amount of irrigated land (Bilen, 1994, Table 1). Satellite data can be used to monitor the amount of irrigated land, but with a number of limitations. First, crops which are primarily rainfed, but with supplemental irrigation in dry years will be difficult to identify. Second, the annual sequence of vegetation is highly variable in irrigated land. Fields may produce one, two or even three crops each year, unrelated to the timing of wet and dry seasons. Thus they do not have a uniform temporal NDVI signature. Finally, irrigated farming typically has a range of physical scales. A single large scale irrigation project is composed of thousands of much smaller fields, each with a different crop/fallow cycle.

To illustrate the broad distribution of irrigated lands in the Near East, we use a dry season NDVI from 1 km AVHRR images for August 1995? (Figure 15). In this way, we get an impression of the geographical distribution of irrigated lands. Some important areas include (from west to east):

- along the Ceyhan River near Adana, Turkey
- along the Orontes River in Hatay, Turkey
- aqueduct supplied projects west of Lake Assad and at the junction of the Balikh and the Euphrates rivers in Syria
- along the Syria/Turkey border along the Euphrates, at the head of the Balikh River north of Raqqa (i.e. Harran plain in Turkey) and the Khabur River near Hasakah
- srips along the Euphrates river in Syria
- near Kirkuk, Iraq on the little Zab River
- vast fields in Mesopotamia near Baghdad, Iraq.
- vast fields and marsh near Basra, Iraq

A view of irrigation change is seen by comparing images from different years. In Figure 16, we use 8-km AVHRR data to compare Mesopotamia in 1983 and 1993. Vast new irrigation areas are seen near and south of Baghdad, Iraq (red color). Nearly equal amounts of land have been taken out of irrigation near Basra (blue color). Little change is evident near Kirkuk over this time period. In Figure 17, we zoom into three regions using 1-km AVHRR. Along the Turkey-Syria border, regions along the Euphrates and in the Harran Plain show massive increases between 1986 and 1996; part of the Turkish GAP project. A central regon near 38°30'E show little change. Rapid growth is seen East of Aleppo near the Lake Assad (fig 17b). Near Kirkuk, Iraq, significant growth occurred between 1995 and 1998. To examine change along the Euphrates River near the junction with the Khabur, we zoom in further using Landsat TM images from 1984 to 1990 (Fig 18) . In Zoom #1, most of the growth has occurred in blocky and linear fields, along the southwestern bank, possibly outside of the flood plain (Fig 18b). In Zoom #2, along the Khabur, there has been striking reduction of irrigated agriculture. Zoom #3, southeast of the junction, shows growth in and along the margins of the flood plain.

While these various types of zooms are instructive, we seek more quantitative methods for describing changes in irrigation. One aspect which cannot be ignored is the variety of field sizes within each irrigated "plantation" or flood plain.. Within large areas with available irrigation water, the cycle of yield and fallow is practiced on an array of small adjacent fields. Only Landsat images can resolve these individual fields. An additional problem may be the shifts in the timing of harvest, if we use a fixed observing time to estimate crop activity.

7.2 The size distribution of irrigated fields

An important aspect of the spatial pattern of irrigation is the size distribution of fields. The size distribution may influence the way we estimate the area of irrigation. To consider this issue, the size distribution for three regions was determined by processing an August TM image in the following way. First, a histogram of NDVI values is plotted for a large "plantation". A clear bimodality in this histogram indicates that a threshold value of NDVI can be used to distinguish active from fallow pixels. Second, using this threshold, raster to vector conversion is done (using ArcInfo) creating a list of N active fields (Figure 19), each with area A_i. Field area range from the size of one TM pixel (about 0.08 *ha*) to about 500 *ha* (i.e. 5 square km). Using the set of A_i values, the total irrigated area is

$$TA = \sum_{i=1}^{N} A_i \tag{8}$$

Third, the N fields are assigned to one of M size categories. Each size category has a width (i.e. bin size) $dA_j = A_{max} - A_{min}$, a mean area A_j and a number of fields N_j. In this representation, the total irrigated area is

$$TA = \sum_{j=1}^{M} N_j A_j \tag{9}$$

From the N_j values, the size distribution function $f_j = N_j / dA_j$ can be defined with units ha^{-1}. This definition avoids the effect of variable bin size and allows us to take the limit as $dA_j \to 0$. The total irrigated area becomes

$$TA = \sum_{j=1}^{M} f_j A_j dA_j \tag{10}$$

or, as $dA_j \to 0$,

$$TA = \int_{A\min}^{A\max} f(A) A dA \tag{11}$$

A log-log plot of $f(A)$ for region 1 is shown in Figure 19, indicating that $f(A)$ may be approximately of the form

$$f(A) = CA^n \tag{12}$$

so that

$$\log f(A) = \log C + n \times \log A \tag{13}$$

Whereupon, from (11, 12)

$$TA = \int_{A\min}^{A\max} CA^{n+1} dA = \frac{C}{n+2}\left[A^{n+2}\right]_{A\min}^{A\max} \tag{14}$$

For region 1, 2 and 3, the values of C are C=1763, 2618 and 2013 respectively while the exponents are n = –1.83, -1.923 and -1.87. These coefficients are computed with a least-square fit with *R*-squared values 0.92, 0.94 and 0.90. In the fitting process, the smallest fields are truncated as they are not well measured and the few largest fields are truncated as being statistically insignificant. The *C* values reflect the overall size of the plantation and active field density within, while the n values are characteristic of the nature of the farming practices and land ownership. We test the accuracy of a distribution function representation by comparing the total areas computed from (14) using A_{min} = 1.35 ha and A_{max} = 56.2 ha, with actual areas from (8). The results are (9644, 11,570 and 10,030 *ha*) , while the actual total areas from (8) are (17,270, 15,700 and 13,500 *ha*). The former values are underestimates, primarily because of the truncation of large and small large fields.

The spread in n-values for regions 1, 2 and 3 is quite small, indicating perhaps a universal nature to this type of size distribution. With *n* near –1.9, the exponent in the integrand of (14) (i.e. $n+1 = -0.9$) is negative indicating that smaller fields will make significant contributions to the total area due to their greater number. If the actual evaporation (*AET*) from irrigated fields is strictly proportional to the total area, then *AET* too will have a significant contribution from the smaller fields.

According to some physical models of evaporation, the growth of an "inner boundary layer" downstream of the windward field boundary will decrease the water vapor flux as the square root of the fetch. Such a model would yield a non-linear relationship between field area and *AET*: i.e. $AET \sim A^{(3/4)}$ for square fields. In this case, the total *AET* is proportional to

$$AET \sim \int_{A\min}^{A\max} f(A)\, A^{3/4} dA = \int_{A\min}^{A\max} C A^{n+3/4} dA = \frac{C}{n + 7/4} \left[A^{n+7/4} \right]_{A\min}^{A\max} \qquad (15)$$

In (15), with $n = -1.9$, the exponent $n+3/4 = -1.15$ is negative, indicating that smaller fields will make a larger contribution to the total *AET*, due to the greater number of these fields and the effect of their isolated nature on the inner boundary layer. The uncertainty in *AET* introduced by the broad range of field sizes, not to mention the influence of field adjacency (not treated here), poses a challenge to the estimation of *AET* from space which goes beyond the scope of this study.

7.3 Monitoring the total irrigated area

The primary difficulty in large scale temporal monitoring of irrigated area is related to the wide range of field sizes and the significant contribution from small fields. As Landsat images, which can resolve small fields, are available infrequently, we must rely on AVHRR for continuous monitoring. Unfortunately, the AVHRR pixel size of 1 km^2 (100 *ha*) is larger than all but the largest irrigated fields. The 8×8 *km* spatial resolution in some of the Composite AVHRR imagery is even less capable of resolving individual fields. It follows that any direct method of area estimation using AVHRR will encounter a serious "mixed pixel" problem, i.e. each pixel will contain both active and fallow or non-irrigated fields. When only a small percentage of the pixel area is active, the average NDVI for the pixel will be very sensitive to extraneous factors such as haze, sun angle, sensor drift, sparse surrounding vegetation and so on. Setting a threshold value for pixel NDVI to indicate irrigation leads typically to considerable overestimates and wild interannual variation in total area.

As a solution to this problem, we use a statistical downscaling technique. Using simultaneous Landsat TM and AVHRR images for the month of August, a linear regression relationship is determined between irrigated area (A) in 1-km or 8-km Landsat TM blocks and the corresponding pixel NDVI value in the AVHRR image. This relationship is of the form

$$A = b \times (NDVI - NDVI_{ref}) \qquad (16)$$

where NDVI varies between -1 and +1. In (16) the reference NDVI is taken from a large region of adjacent desert or steppe with no irrigation. The reference NDVI value may vary from year to year due sensor drift or other effects. According to (16), the amount by which NDVI exceeds the reference value, is proportional to the percentage of irrigated land within the pixel.
Values for b are different for different data sets (i.e. 2.345 for 1km data and 1.032 for 8-km) as their processing and scaling by the USGS are different. For example, if a 1-km pixel has NDVI = 0.3 and the reference value is 0.1, the proportion of irrigated land is 0.46 and amount of irrigated land in that pixel is 0.46 km^2 or 46 ha. By using a smooth function like (16) instead of a threshold, and including a reference value, a slight drift of NDVI due to extraneous factors will produce only slight changes in the area estimate.

The time series of irrigated area deduced from the proportional method (16) are shown in Figure 21 and Table 6. Twenty three regions are shown, exhibiting a variety of histories. To describe these results, we start with the areas in Turkey and work our way southeast along the Euphrates into Mesopotamia.

The time series for four irrigated regions in Turkey are shown in Fig 21a. A region near Adana, fed by the Ceyhan River, has grown steadily from about 750 km^2 in 1982 to 1400 km^2 in 1998. Near Antakya in Hatay, the irrigated area was constant at about 550 km^2 until 1994; then grew rapidly to 1050 km^2. West of the Harran Plain, the area has fluctuated about a value of 100 km^2, with little net change. In the Harran Plain, irrigated area grew from to 90 km^2 in 1982 to 400 km^2 in 1994 followed by a rapid rise to 1200 km^2 in 1998. This growth was made possible using water from the new Ataturk dam at 37.5°N and 38.5°E.

Along the Syria/Turkey border, irrigated area along the Euphrates River (north of Lake Assad) grew slowly at first and later more quickly from 70 km^2 in 1982 to 200 km^2 in 1998 (Figure 21b) . In Syria, just south of the Turkish border near 40 to 41°E, area A has fluctuated about a constant area of 100 km2 while area B increased slowly from 25 km^2 in 1982 to 150 km^2 in 1994; and then rapidly to 400 km^2 in 1998. Just west of Lake Assad (east of Aleppo) , irrigated area increased from 20 km^2 in 1982 to about 100 km^2 in 1992; followed by a rapid increase to 500 km^2 in 1998.

We now consider the irrigated regions along the Euphrates River downstream from Lake Assad (Figure 21c). The region along the Balikh north of Raqqa, rose from 40 km^2 in 1982 to 100 km^2 in 1992; followed by a slightly more rapid rise to 250 km^2 in 1998. The Euphrates SE of Raqqa rose smoothly from 180 km^2 in 1982 to about 500 km^2 in 1998. The Euphrates NW of Deir Ezzor rose smoothly from 80 km^2 in 1982 to 270 km^2 in 1998. SE of Deir Ezzor, irrigated area along the Euphrates rose smoothly from 200 km^2 to 400 km^2.

Irrigation along the Khabur is plotted in Figure 21d. Area C (south of the Turkish border) increased from 200 km2 to about 500 km2 between 1982 and 1998. The upper Khabur increased fro 100 km2 to 200 km2. The central Khabur increased from 40 km2 to 80 km2. The lower Khabur has reamined constant overall, perhaps limited by increased water consumption upstream

Finally we consider irrigation in Iraq (Figure 21d, e). The area near Kirkuk, fed by the Little Zab, rose from near zero in 1982 to about 250 km^2 in 1992; followed by a more rapid rise to 1000 km^2 in 1998. Massive increases in area are seen in Mesopotamia near Baghdad. The area north of Baghdad rose from 600 km^2 in 1982 to 1000 km^2 in 1992; followed by a rapid rise to 2200 km^2 in 1998. A similar profile is seen just east of Baghdad. South of Baghdad, area was constant at 400 km^2 until 1992, then rose to 1000 km^2 in 1998. The large region south east of Baghdad showed a similar accelerating increase, but with a decline since 1996. The increased rate of growth after 1992 in Iraq is probably government policy associated with the Gulf War and subsequent embargo.

The region near Basra is also of interest (Figure 21e). The large marsh area west of Basra , the Shatt el-Arab, exhibits a brief dip in vegetation in 1984 and a sudden collapse from 1300 km^2 to near zero in 1992-1994. The dip and collapse may be a draining of the marsh area, perhaps driven by military and political decisions connected with the Iran-Iraq war and the Gulf war. Water resources may also be at issue considering the amount of flood control and irrigation development which has occurred upstream on the Euphrates (Beaumont, 1998) . The area east of Basra, along the border with Iran, has remained constant at about 450 km^2.

The information on irrigated area is summarized in Table 7. Large increases are seen in all three countries: Turkey, Syria and Iraq. The increases clearly accelerate in the early 1990's, probably due to a combination of capital availability, marketplace globalization and population increase. The impact on water resources is discussed below (see also Beaumont, 1998).

[Table 6]

Estimates of water usage in intensive summer irigation can be estimated as follows. Adding the values in Table 6 for the Upper Euphrates (i.e. everything in Figure 15a except Adana and Hatay) the total irrigated area in the basin in 1998 to be about 4,000 square kilometers. At an August temperature of 30°*C* the potential evaporation is $PET=5\times30= 150\ mm/month$. Assuming that $AET=PET$ because of the availability of surface water, the August evaporation from irrigated land is

$$AET \times A = (.15m)\left(0.4\times10^{10}\,m^2\right) = 600\ mcm \qquad (17)$$

From Table 5, the average annual Euphrates discharge is about 27,000 mcm. The evaporative loss (17) is about 2% of this value. The loss (17) is roughly equivalent to the natural August Euphrates discharge of 350 m^3/s or 906 *mcm* (Figure 16), indicating the essential role of water storage to maintain river flow during the dry season. If the total loss is computed for a four month growing season, the water loss will be 2400 mcm ,nearly 8% of the total average Euphrates discharge. Such usage might not be sustainable in a sequence of dry years. As discussed below, annual precipitation can vary by 50% from the climate average and , as seen in Section 6, the river discharge has a doubled sensitivity because of the threshold nature of runoff.

An alternative way to estimate annual water use from irrigated area is to use an empirical value of 10,000 m^3/ha (Beaumont, 1998), equivalent to 1 meter of evaporation. This value is about twice the four month evaporation given by (17). Using 4,000 km2, the annual water use thus computed is 4000 mcm; about 16% of the annual Euphrates discharge. .

A similar calculation can be done for Mesopotamia. Summing the four areas near Baghdad gives about 6000 km^2. Using this area in (17) gives a water use of 900mcm for August and 3600mcm for a four month summer season. This latter figure represents about 13% of the annual average Euphrates discharge. Using the alternative method, this estimate jumps to 26%.

Clearly, Near East irrigated agriculture in 1982 was running far below its water resource capacity; but this situation is now changing. Slow increases in irrigated area in the 1980s followed by rapid increases in the 1990s have brought water demand to a level which might not be sustainable in drought years . An extrapolation of current area growth trends (i.e. 1992 to 1998) for another decade is probably not physically possible, unless AET can be reduced by more efficient water application.

8. Rainfed Agriculture

To examine the sensitivity of rainfed agriculture to interannual climate variation, we start by observing the variability in NDVI near the end of the winter/spring growing season. In Figure (16) we present the map of normalized standard deviation (i.e. the coefficient of variation) for the April/May NDVI. Excepting two irrigated areas in the far NW and SE corners of the region, the highest variability is seen in a crescent-shaped region lying in the steppe; corresponding approximately to the Soil Moisture Class 5, Temperature Classes 6 and 5 and SWAP13/93 Classes 3 and 12. Soil Moisture Class 5 shows a steeply descending soil moisture content during April and May. Temperature Class 5 has an average minimum January temperature of +3C and thus will experience brief frosts, the severity of which may vary from year to year. SWAP13 Class 3 and 12 were identified as May-peaking vegetation classes; mostly rainfed grain agriculture.

In order to objectively assess the role of climate variation, and distinguish between different types of climate sensitivity, we define a number of annual climate indices (C_i), computed by averaging together data from six climate stations in NW and in NE Syria. From these stations, we compute average precipitation and maximum, minimum and average temperature for the Fall (October/November), Winter (Dec/Jan/Feb), Early Spring (Feb/Mar April) and Spring (Mar/April/May); 16 indices in all;, 32 if the two station clusters are used. For the period with Pathfinder satellite coverage, (1982 to 1993) each climate index is then correlated with the April, May and June NDVI (N_i); on a pixel-by-pixel basis. The correlation coefficient between climate index and NDVI is calculated using

$$R = \sum_{i=1}^{12} (N_i - \overline{N})(C_i - \overline{C}) \times (\sum_{i=1}^{12} (N_i - N)^2 \times \sum_{i=1}^{12} (C_i - \overline{C})^2)^{-1/2} \qquad (18)$$

To condense all this information, we select the three climate indices that correlate most strongly and broadly with spring vegetation: these are winter minimum temperature and winter and early spring precipitation. All three of these correlation coefficients are positive in our domain. In Figure 23, the pixels for which a correlation coefficient exceeds R=0.45 are assigned to one of three classes, depending on which of the three climate indices gives the highest correlation. This moderate threshold avoids most of the accidental correlation and the choice of the best correlating index avoids most of the secondary correlations. Pixels with R<0.45 are left unclassified. Other methods of climate-vegetation correlation analysis are described by Kogan (1997).

The three spring months show distinctive patterns of vegetation sensitivity to climate. In April (Fig. 23a), three regions are clearly seen. In the southeastern Turkey and northwestern Iran NDVI has a significant positive correlation with the minimum winter temperature. In northeastern Syria and into Iraq, the NDVI correlates well with winter precipitation. In the vast steppe area between 33 to 35°N and 37 to 43°E, NDVI correlates positively with spring precipitation. Our physical interpretations of these three types of correlation are:

- Low winter minimum temperatures induce frost damage or delay root growth in winter wheat
- High winter precipitation allows water to be stored in the soil improving the spring growth of winter wheat and barley
- High spring precipitation improves the greening of annual grasses in the steppe.

The regions in Figure 23a that are sensitive to interannual fluctuations in winter minimum temperature correspond most closely to temperature class 5 (Fig 2) . In Figure 3, temperature class 5 has a mean January temperature of 3°C; thus an abnormally cold year could lead to extensive frost damage. The regions in Figure 23a that are sensitive to interannual fluctuations in precipitation correspond most closely to precipitation classes 3 and 4 (Figure 4) and soil moisture class 5. As seen in Figure 6a, the number of wetdays (WD= 125 to 150days) in this climate zone is quite sensitive to annual precipitation.

In May (Figure 23b), the climate sensitivity patterns are somewhat similar. The winter-minimum-temperature-sensitive belt has shifted slightly northward. The winter-precipitation-sensitive belt in northeastern Syria has weakened but new areas in the mountains of Iraq have appeared. The spring precipitation-sensitive belt has changed its shape, mostly by losing the Jazirah and shifting northward.

In June (Figure 3c) the pattern is dramatically different. Little sensitivity to winter minimum temperature or spring precipitation remains. The region sensitive to winter precipitation is vast, extending diagonally from 38N/40°E to 33N/50°E. The northward shift in this region, from April to May to June, is due to the fact that the wheat matures and is harvested later in the higher elevation, more northern latitude and cooler climatic areas.

One other feature in these figures deserves comment. Note that the regions of traditional irrigation (e.g. along the Euphrates) show little correlation with climate. Established irrigated farming should not be sensitive to precipitation and, as they lie in the warmer zones (T Classes 7, 6), frosts are not common. The signs of the R-values (both positive) are as expected. Increased winter precipitation, stored in the soil until Spring, should improve the maturation of wheat and barley. Warmer winter temperature should prevent strong winter frosts which could stunt or injure early shoots.

Two uncertainties may confuse our interpretations. First, our time series is very short (12 years) and thus accidental correlations may occur. The fact that we have looked at several dozen index combinations raises the probability of such an accident. The spatial continuity of the correlation maps is reassuring on this point, as is the fact that both climate station clusters give similar results. The second uncertainty is the possibility of secondary correlations. If the different climate indices are correlated among themselves, then a physically-based NDVI correlation with one index will promote unphysical correlations with others. As indicated above, we accept as meaningful, only the single climate index with the highest correlation coefficient with vegetation. In addition, we note that the NDVI-climate correlations computer using an independent set of climate stations in northeastern Syria show very similar results (not shown).

The three climate sensitivity maps (23-25) provide a new view of how the landscape responds to interannual climate variation. The Middle East is a mosaic of different climate-sensitive regions.

9. Summary of landscape sensitivities

In the previous four sections, we have attempted to quantify the response of landscape and water resources to climate variability and human action. We begin this section by summarizing these response functions; especially the impact of temperature, precipitation and changes in agricultural policy.

Increases in temperature will:

- Decrease the winter snow cover in the Tauros and Zagros mountain study area by about 50,000 km^2 per degree Celsius. The snow line will rise about 200 meters per degree Celsius.
- Due to increased evaporation, the annual discharge in the Euphrates and Tigris will decrease by about 2000 *mcm* and 4000 *mcm* per degree Celsius respectively
- Due to reduced snow storage, the spring peak in Euphrates and Tigris runoff will be reduced and shifted earlier by about one week per degree of warming. Warming exceeding 5 degrees Celsius may smooth out and reduce the spring discharge peak by nearly eliminating the role of snowpack storage.
- In the central part of the Fertile Crescent (E38 to E46), due to the steep orographically-amplified N-S temperature gradient, the winter frostline will move northward less than 20 *km* per degree Celsius.
- Due to the weak E-W temperature gradient between N30 and N37, vast new areas, especially in Syria and Jordan will move into the zone with summertime temperatures exceeding 30°*C*
- Due to the recent growth of irrigated area, now averaging 15,000 km^2 in Syria, Turkey and Iraq, the JJA evaporative loss will increase by about 140 *mcm* per degree Celsius in each country.
- In the Fertile Crescent, due to increased evaporation, the moist growing season (i.e. Wetdays) will decrease by about 20 days per degree Celsius. This will move the boundary between desert and steppe northward about 100 *km* per degree Celsius.
- Spring leafing-out or snowmelt in upland areas may occur earlier.

Increases in precipitation will:

- Delay the springtime final melting of the snowpack on the mountain slopes. This delay is evident from satellite images but causes no significant change in the shape of the modeled run-off profile
- Increase annual river runoff in direct relation to the increased precipitation
- Increase the moist growing season (i.e. wetdays) by about 15 days for each extra cm of precipitation. This moves the boundary between desert and steppe about 75 *km* southward.
- Strongly improve the success of marginal barley farming on a 150 *km* wide belt in the steppe
- Increase the spring greening of annual grasses in the dryer steppe.

The "satellite-evident" changes in agricultural policy are:

- Rapid sustained increased in irrigation, between 1982 and the present, in new aqueduct-fed plantations, along the traditional river-courses and in broad areas of Mesopotamia. The summer JJA evaporation from these fields now approaches the annual runoff in the major rivers.

While previous standard analyses have identified most of these changes, our satellite and model based approach adds a quantitative aspect that allows us to consider various climate change scenarios. We are particularly interested in scenarios which include couple changes in temperature (dT) and precipitation (dP). We consider the following scenarios: interannual fluctuations, thermodynamic and speculations of climate change

10. Climate change scenarios for rainfed and irrigated agriculture

10.1 Coupled temperature and precipitation changes

There is a good deal of interest in climate change in the Middle East. Two particular periods of interest are the Third Millenium BC in relation to the history of civilization and agriculture and the 20th and 21st Century in relation to recent and future greenhouse warming. Convincing methods for analysing altered climates are just emerging. Global models (i.e. Kutzbach, 1996) have limited application in small regions whose climate is dominated by local water bodies and terrain. To overcome this problem, several new methods have been proposed for "downscaling" , that is, transforming information about large scale changes to smaller scales (Georgi and Mearns ,1991). To date, these methods have not been applied to the Middle East region, but some insight can be gained by reviewing recent work in Alpine Europe. The Alps are roughly similar to Tauros and Zagros in their geology, terrain height, climate and in the central role they play in the river hydrology of their region. Luthi et al (1996) have tested a regional model for its ability to capture interannual variation. Schar et al. (1996) used that regional model to understand the response of Alpine precipitation to 2°C warming with unchanged relative humidity. Neglecting any changes in large scale wind patterns, they found that the corresponding 16% increase in absolute humidity resulted in only a 4% increase in precipitation in central and Alpine Europe. An alternative method is described by Burkhardt (1999) in which the $3\times CO_2$ global climate model (GCM) predictions of altered 500hPa geopotential fields (and related wind fields) are statistically linked to altered precipitation patterns. Burkhardt neglected temperature and humidity changes. As the modified global climate had stronger southerly winds over Europe, their model predicts increased/decreased precipitation south/north of the Alps. Typical changes are less than 10%. A method of nesting the Alpine region into a GCM is described by Jones et al (1997). Their method takes into account both circulation and humidity changes.

The inconsistent results from these different techniques indicates that even these rather advanced methods may be unreliable. Confidence in regional climate prediction will require improvements in large scale wind field predictions and in the physics of mesoscale precipitation processes.

Given the limitations on regional climate modeling and the fact that no nested Middle East modeling has been carried out, we restrict our inquiry to one question: "Is rainfed and irrigated agriculture more sensitive to climatic variation in rainfall or temperature?" This question is a useful one as it forces us to think of climate change as a coupled change in precipitation and temperature. We proceed by considering four climate change scenarios; i.e. specified coupling of T and P changes, defined by the ratio (S) of proportional precipitation change ($dlnP = dP/P$) to temperature change (dT).

$$S = dlnP/dT \qquad (18)$$

with units C^{-1}. These scenarios are summarized in Table 6.

[Table 7]

. The S1 scenario is based upon the type of data shown in Fig 22. Interannual variability in the Fertile Crescent is characterized by a DJF standard deviation of about 1.5°*C* in temperature and about 1.5 *cm* out of a mean 3.5 *cm* in precipitation. The proportional deviation in precipitation is therfore $dlnP$ = 1.5/3.5 = 0.43. The ratio $S1$= 0.34/1.5 = 0.29 C^{-1} We can consider this ratio to represent a "scenario", even though the two changes might not be causally or statistically related.

The S2 thermodynamic relationship between P and T changes is the Clausius-Clapeyron relation for the saturation vapor pressure (e_s)

$$\frac{1}{e_s}\, de_s = \frac{\left(\frac{L}{R_v}\right) dT}{T^2} \tag{19}$$

where L and R_v are the latent heat of evaporation (2.5×10^6 $J\,kg^{-1}$) and the gas constant for water vapor (461 $J\,C^{-1}\,kg^{-1}$) (Wallace and Hobbs, 1977). If the relative humidity of the atmosphere remains constant, along with all atmospheric dynamical and microphysical processes, the precipitation will increase in proportion to e_s . If T=273K , and for a temperature change dT (19) gives

$$\frac{1}{P}\, dP = 0.0728\, dT \tag{20}$$

giving S2=0.73. Thus a one degree rise in temperature would increase e_s and P by 7.3%. This thermodynamic relationship was the input to the Alpine climate change models of Schar et al. (1996) and Frei et al. (1998).

The other two scenarios (S3 and S4) derive from Schar et al (1996) and Burkhardt (1999). Both of these studies find rather small precipitation changes in relation to temperature changes.

10.2 River discharge and Irrigated agriculture

In Section 7, we saw that August irrigated agriculture in the steppe and desert is decoupled from local precipitation but dependent on river water. We saw in Sections 3 and 5 that the storage of upland snow and the timing of snowmelt and river discharge is governed primarily by temperature. The total annual discharge (D) however is sensitive to both temperature and precipitation. Consider the Euphrates River. From Table 5, the sensitivity factors can be computed: $dD/dT = 22\times10^3$ *mcm*/10°*C* and $dD/dlnP = 25\times10^3$ *mcm*/0.5. For small variations, these two influences can be added to give a general climate response formula

$$dD = -2.2\, dT + 50\, dlnP \tag{21}$$

in units of 10^3 *mcm*. Thus a warming of one degree, or a precipitation increase of one percent, would decrease D by 2200 *mcm* or increase D by 500 *mcm* respectively.

The relative importance of the temperature and changes in (21) is determined by the coefficients of the two terms and ratio S in Table 6. For interannual variation (S1 = 2.9) , the precipitation will dominate. For example, if we enter the interannual standard deviations of T and P into (21), we obtain

$$dD = -2.2\ (1.5) + 50\ (0.43) = -3.3 + 21.5 = 18.2 \times 10^3\ mcm \tag{22}$$

The thermodynamic scenario *S2* has a smaller ratio *S2*=0.073. Putting this value into (21)

$$dD = (-2.2 + 50 \times 0.073)\ dT = (-2.2 + 3.5) dT = +1.3\ dT \tag{23}$$

Thus, a temperature increase of one degree, with an accompanying thermodynamic precipitation increase will increase the Euphrates discharge by 1.3×10^3 *mcm*. This rather small increase could be cancelled by the increase irrigation of water needed to balance crop evapotranspiration.

We conclude that if climate change mimics interannual variation, the changes in precipitation are likely to be more significant than temperature changes. If on the other hand climate change follows a simple thermodynamic scenario, changes in precipitation and evaporation will nearly cancel and the net effect on annual river discharge will probably be quite small. By extension, scenarios S3 and S4 will give a reduction in discharge due a dominance of evaporation.

As shown in Section 7, the immediate future of irrigation will be dominated by new irrigation projects. For example, an increase of summer irrigation area along the Euphrates of 15,000 km^2, with a JJA temperature of 30°C will increase the required water by

$$AET \times Area = (.45\ m)\ (.5 \times 10^{10}\ m^2) = 6750\ mcm \tag{24}$$

This increase is about 40% of the interannual variation in D given by (22). When this demand is added to the corresponding increased demand for supplemental irrigation, the river discharge in dry years will be insufficient.

10.3 Rainfed agriculture

Our method for estimating the relative effect of *T* and *P* on rainfed agriculture is based on the calculation of the length of the moist season described in Section 3. As the number of wet days is related simply to the Koppen-Geiger index ($K=P/T$), we need only consider changes in this quantity (dK).

$$dK = \frac{dP}{T} - \frac{PdT}{T^2} \tag{25}$$

Once again, the relative values of *dP* and *dT* determine whether the change in precipitation or evaporation will dominate. Using the interannual standard deviations and means; dT=1.5°*C*, T=15°*C* and dP=1.5 *cm* and P=3.5 *cm* , (25) gives

$$dK = (1.5cm)(15°C)^{-1} - (3.5cm)(1.5C)(15°C)^{-2} = 0.1 - 0.02 = +0.08\ cm\ °C^{-1} \tag{26}$$

Such an increase in K could cause a substantial increase in the length of the moist growing season (Figure 6b). The dominance of precipitation over temperature is consistent with the strong correlation found in Section 8 between NDVI and precipitation. No consistent correlation between NVDI and T was found in water stressed areas.

A parallel calculation can be done using the thermodynamic *S2* scenario. Using (22), if dT=1*C*, the corresponding $dlnP$ = 0.073. If P=3.5 *cm*, $dP = 0.073 \times 3.5 = 0.26$ *cm*. Substituting these values into (25),

$$dK = 0.26cm/15C - (3.5cm)(1.0C)/(15C)^2 = 0.017 - 0.015 = +0.002\ cm\ °C^{-1} \qquad (27)$$

Note that the two terms nearly cancel. As seen in Figure 6b, such a small change in K will barely alter the length of the moist growing season. Scenarios S3 and S4 will reduce the KGMI index.

One significant exception to the dominance of precipitation must be remembered. According to Figures 23, 24, the northern edge of the Fertile Crescent is rather sensitive to winter minimum temperature. This region responds to frost instead of soil moisture variations. A warming of the climate would aid agriculture in this region, independent of the trend in precipitation. Further south, in the steppe, another direct effect of temperature might be felt. Heat stress on plants could increase if climate warming pushes spring temperatures toward 30°C. This influence was not detected by our satellite correlation method however.

11. Summary

In this paper, we have tried to compactly describe the large scale aspects of the changing landscape of the Middle East, using climate and satellite radiance data and a simple water budget model. The results indicate that the region is a mosaic of differing climate sensitivities. The high mountains, the farmland of the Fertile Crescent and the steppe each respond to different climate indices, with different time lags. These sensitivities have implication for agriculture in the region. A warmer climate will improve agriculture in the northern part of the Fertile Crescent while drying out the southern part. If the warming is accompanied by slightly increased precipitation (i.e. the thermodynamic scenario), drying of the southern part will be insignificant. If the warming brings larger gains in precipitation, soil moisture could improve overall.

Similar results are found for the river discharge that supplies irrigated agriculture. Warming would decrease water storage in mountain snow, and decrease annual discharge totals. An accompanying increase in precipitation could allow the annual total discharge to maintain its level or increase. However, recent rapid increases in irrigated area have brought water usage close to supply levels: a trend which dominates any conceivable change in climate.

An auxiliary objective of this study is to provide a framework for future studies of the Middle East landscape. A few specific recommendation are listed below.

- Land Cover/Land Use analysis should include interannual variability. Sensitivity to climate is an important characteristic of Land Cover.
- Higher quality satellite images are essential for Land Cover analysis in arid regions. Better radiometric resolution and calibration, compositing, atmospheric correction, bidirectional reflectance and spectral band selection will allow much more precise analysis.
- The standard method of multitemporal NDVI LC/LU classification must be extended to avoid confusion between distinct classes; especially confusion with irrigated agriculture. Fourier time series analysis and additional data types should be used.
- The statistical distribution of agriculture field sizes requires a new approach to the spatial analysis of evaporation.
- The complexity of hydrologic modeling should be minimized, to correspond to the analysis objectives, the quality and type of input data and the potential for verification. Simple models promote robustness and reproducibility.
- Estimating the relative impact of changing temperature and precipitation in the Middle East requires both a landscape climate sensitivity analysis, as given here, and a dynamic model of circulation changes in an altered climate. Neither interannual variation nor a thermodynamic scenario are good proxies climate change.

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Appendices

1. Data sources
2. The hydrology model

Appendix 1. Data sources

Several types of data are incorporated into this work. The climate data available to us includes the FAO 25-year averaged climate station data and NCAR Summary-of-the-Month monthly data from a smaller group of stations. The region covered by our analysis, and the location of FAO stations, are shown in Figure 1. Neither of these climate data sets are fully suitable for the task at hand, but they provide a reasonable first approximation to the average climate and interannual variability of the SWAP region. A digital topography data set for the region was obtained from the thirty-arc-second global DEM at the USGS. Satellite data include two types: AVHRR (poor spatial resolution; good temporal resolution) and Landsat (good spatial resolution; poor temporal resolution). Three types of AVHRR data were used. First, the 8 *km* Composite Pathfinder AVHRR data set provided continuous monthly coverage from 1982 until 1994. Second, the 1 *km* Composite AVHRR data, for ten-day periods during 1993 and 1995, formed the basis for our vegetation class map. Third, AVHRR “snapshots” from the recent period (1993 to 1998) were used to bring our irrigation analyses up to the present day. The Landsat data included both MSS data (the mid-1980s) and the more capable TM data (the 1990s). Discussions of the quality and use of AVHRR data are given by James and Kalluri (1994) and Holben (1986).

In the future, analyses of the type described here should be done with more detailed conventional climate data sets and satellite data from the more accurate, stable and higher resolution sensors on the next generation of satellites.

The data types used in this analysis follows:

- FAO climate station data (1955-1972)
- NCAR Summary of the Month climate station data
- USGS/NOAA Monthly 8 *km* Composite AVHRR (1982-1993)
- USGS/NOAA 10-day 1 *km* Composite AVHRR (1992-3 and 1995)
- NOAA 1 *km* AVHRR
- USGS Dtopo30 Digital Elevation Model
- Digital Chart of the World political boundaries (modified)
- USGS watercourses, seas and lakes
- Ground truth expeditions to Syria (Dec. 1997, April and Nov. 1998)
- Landsat MSS and TM imagery
- River discharge data from various sources

Appendix 2. The hydrology model

In constructing a hydrology model for the Middle East, we are confronted with an enormously complex landscape and very little input data. The only climate data with regional coverage is monthly temperature and precipitation data from about one hundred stations. No reliable wind, sunlight or humidity are available. The terrain ranges from flat low basins to steep high mountains. The temperature and precipitation have strong seasonal and spatial variations.

Mindful of the inconsistency of using a complex model with limited input data (Jakeman, Evans) , we have chosen a simple linear model which includes the three most important hydrological variables; temperature, precipitation and terrain height and roughness. This scheme is basically a "bucket" model (Thornthwaite,) which maintains mass budgets for soil moisture and snow. This model allows computation by hand calculator for individual pixels and, with simple inputs, closed form solutions can be derived. For general inputs, we solve the model equations analytically during each month. Subsequent months use the end values from the previous month as initial conditions. A sequence of such calculations gfives the seasons cycle.

The first step in the analysis of soil moisture, runoff and snow is the interpolation of monthly temperature and precipitation values to a uniform 5 *km* grid. This interpolation is accomplished with a Gaussian weighting function $W=exp(-X)$ with $X=(r/R_0)^2$ where r is the horizontal distance from a grid point to a climate station and R_0 is a variable radius of influence. The value assigned to each grid point is then

$$T\left(x,y\right)=\frac{\sum W\times T_i}{\sum W} \qquad (1)$$

summed over all stations in the vicinity of the grid point. A key problem in the interpolation of climate data is the variable density of stations. We account for this by allowing the radius of influence to vary from grid point to grid point. For each grid point we start with R_0=25 *km* and increase it until the sum of weights ΣW=1.1. This choice insures that each assigned grid value with take into account data from at least two stations. This iterative method is computer intensive but gives smooth fields without spurious extrema and captures the finest structure allowed by the local station density.

The interpolated temperature field is corrected for altitude according to

$$T_g = T_I + \gamma \times z \qquad (2)$$

where T_g is the assigned grid point temperature, T_I is the interpolated sea level temperature, γ is the assumed lapse rate and z is the grid point altitude above sea level. The chosen lapse rate (γ= -5º*C/km*) is similar to values used in studies elsewhere. It was verified locally by plotting the lapse rate (*dT/dz*) between every station pair as a function of horizontal distance between the stations. The resulting scatter plot (not shown) converged nicely to –5 +/- 0.5º*C/km* for small interstation distance. The lapse rate correction plays a significant role in generating small scale temperature patterns unresolvable by the coarse (and elevation biased) station network.

The same interpolation scheme is used for precipitation, but no correction for altitude is applied. The recent literature suggests that simple altitude corrections fail to capture the sheltering of one mountain range by another and in the best case are valid only in small regions. Thus, they would introduce additional uncertainty in our results. In future work, we plan to construct a physical model of precipitation patterns in the Middle East.

The assumptions and equations of the model are given below.

1) The maximum soil moisture is (*MSM*) is assumed to be uniform over the entire region, independent of soil thickness or type. This parameter is used primarily to set

evapotranspiration, but it also plays a role in runoff, as described below. In standard runs, *MSM* is taken to be 100 *mm* of column equivalent water (i.e. 10 grams of water per square centimeter).

2) The monthly Potential Evapotranspiration (*PET*) is controlled by the monthly averaged air temperature (*T*), the number of days in the month (*M*) and the day length (*D*) according to

$$PET = 5.2 \times (T - T_{crit}) \times \left(\frac{M}{30}\right) \times \left(\frac{D}{12}\right) \quad for\ T > T_{crit} \qquad (3)$$

and

$$PET = 0 \qquad for\ T < T_{crit} \qquad (4)$$

where *PET* is expressed in mm/month. The linear relationship between *T* and *PET* in (1) is a simplification of the two-part Thornthwaite (1948) formula. The day length is given by

$$D = \frac{24}{180} \times \cos^{-1}\left(-\left(\tan\phi \times \tan\delta\right)\right) \qquad (5)$$

where δ=23.5 $\times$ sin (d×(360/365)) and ϕ is the latitude of the gridpoint. The quantity "*d*" is the day of the year counted forward from the vernal equinox (March 21). These formulae are illustrated by the following example. Consider a site with a latitude of 35°N and a June average temperature of 35°*C*. According to (4) for June 15 (*d*=91) the day length is about 14 hours. Then with *M*=30, *PET* = 5.2×35×1×(14/12)= 210 *mm/month*. This formula has been checked by comparing it against pan evaporation data from the region. The pan evaporation is about 30% higher than *PET* from (1) as it should be given the isolated nature of the pan device. A choice of T_{crit} =-2.5°*C* gives the right intercept.

3) The Actual Evapotranspiration (*AET*) is assumed to be proportional to the *PET* and the available soil moisture according to

$$AET = PET \times \frac{SM}{MSM} \qquad for\ MSM > SM > 0 \qquad (6)$$

Equation (5) indicates that the *AET* will decrease as the soil becomes drier. This proportional behavior results in an exponential decrease in *SM* following a step input rather than a linear descent to *SM*=0 as in an *SM*-independent *AET* formulation.). Generally, proportional models are preferred in dry climates (Mintz and Walker, 1993, Rutter, 1975). We have also neglected the effect of plant cover on *AET* (see Alen, 1990, Gutman and Rukhovetz,1996 and Moran et al. 1996).

4) The Runoff is assumed proportional to the roughness of the terrain (Sz) and the amount of soil water (SM) according to

$$Runoff = \alpha \cdot Sz \cdot SM \qquad (7)$$

where the coefficient α has units (meter*month) $^{-1}$. 1The roughness for a 5 kilometer grid cell is computed from the 25 one-kilometer cells of which it is composed, according to

$$Sz = (1/25)\sum_{i=1}^{25} | z_i - \bar{z} | \qquad (8)$$

This quantity varies greatly over our domain, from about 2 to 500 meters. Generally speaking, roughness correlates with altitude, but there are exceptions such as high plateaus or basins (i.e. high altitude, small roughness) , and coastal mountain ranges (i.e. low altitude, large roughness).

We assume that water may be stored in the pixel where it precipitated, but upon runoff, it appears instantly in river discharge. Thus, the only storage occurs locally, in the original pixel. Once the water leaves its original pixel we do not track it. It is not allowed to accumulate in another region. Thus, our model is not a true spatial runoff model.

5) The monthly precipitation (*R*), snowmelt (M) and potential evaporation (*PET*) amounts are assumed to be uniform during a month. In this case, we can write a first order differential equation for the time development of soil moisture (*SM*(*t*)) during the month.

$$\frac{dSM}{dt} = R + M - F \times SM \quad (-Excess) \qquad (9)$$

where

$$F(T, Sz) = (PET / MSM + \alpha \cdot Sz) \qquad (10)$$

is the combined rate factor for evaporation and runoff. The term "excess" removes water in excess of the maximum soil moisture (MSM)

In (9) the quantities have been temporarily redefined as rates (units: *mm/sec*) rather than monthly totals. The solution to (6) is

$$SM(t) = SM_{ss} - \left(SM_{ss} - SM(0)\right) \times \exp\left(-F\ t\right) \qquad (11)$$

where *SM*(0) is the soil moisture at the beginning of the month. The mathematical property exhibited in (11) is quite simple. During each month, the value of SM at each pixel relaxes towards it's equilibrium value for that month given by

$$SMss = (R + M) / F \qquad (12)$$

SMss is a function of the monthly averaged temperature and precipitation and the permenent value of roughness (Sz) for the pixel. If the SM value at the beginning of the month is larger (or smaller) than SMss, the value will decay (or grow) exponentially toward SMss.
Evaluating (6) at the end of the month gives

$$SM(T_M) = SM_{ss} - \left(SM_{ss} - SM(0)\right) \times \exp(-F) \qquad (13)$$

where t=T_M at the end of the month and we revert back to *R* and *PET* as monthly totals. If *SM*(*M*) >*MSM*, the excess water is assumed to run off according to Runoff=$SM(T_M)-MSM$.

The amount of runoff and evaporation each month can be computed by integrating (7 , 11)

$$Runoff = \alpha \cdot Sz \cdot ISM + (excess) \qquad (14)$$

And from (6,11)

$$Evap = (PET / MSM) \cdot ISM \qquad (15)$$

Where the monthly integrated soil moisture ISM is

$$ISM = \int SMdt \qquad (16)$$

Which is, from (11), written using monthly values

$$ISM = SMss - ((SMss - SM(0)) / F) \cdot (1 - \exp(-F)) \qquad (17)$$

It is most convenient to think of our parameterizations as selecting characteristic time scales (τ) for evaporation and runoff, because in each case we have set the loss rate proportional to the amount of water in the soil . These time scales however, depend on conditions. The table below exhibits the runoff time scales for various roughnesses found in our domain, with two choices of alpha = 0.02 and 0.05 $(\text{meter*month})^{-1}$

Roughness (Sz)	Runoff timescale (α=0.02/0.05) (τ =1/α Sz)	Terrain description
2meter	25/10 months	Very flat
10	5/2 months	flat
100	0.5/0.2 months	hills
500	0.1/0.04 months (3/1.2 days)	Steep mountains

The table below exhibits the time scales for evaporation, with the choice of MSM = 100mm.

Temperature (T-Toffset)	Evaporation time scale (τ = MSM/PET)
0 C	infinity
10	2 months
20	1 month
30	0.7 months
40	0.5 months (15 days)

It is evident from these tables that cold rough high mountain areas will lose water primarily by runoff while flat hot areas will lose water by evaporation. As seen below, high mountain areas may also store water in the form of snow during the winter, releasing it by melting in the spring. Typically, due to the roughness and low temperature of high terrain, when snow melts it mostly runs off rather evaporates. Excess runoff (SM>MSM) is rare in dry climates. It typically requires SMss>SM (from 12) which in turn requires large precipitation and low roughness.

The combined storage time for water in the soil is

$$\tau = 1 / F \qquad (18)$$

where F is given by (10).

An advantage of our analytical representation of each month's hydrology (11,14,15,17) is that it can handle any time scale value. A numerical solution, even with a short one-day time step, would be inaccurate and could give unphysical negative values of soil moisture. Our method is numerically precise and extremely fast.

5) The amount of accumulated snow is increased by snowfall and decreased by melting, according to.

$$\frac{dSnow}{dt} = P - M \qquad (19)$$

where *Snow* is the water-equivalent snow in millimeters. If the temperature is below Tcrit, the precipitation is assumed to be in the form of snow. If the temperature is above Tcrit, the snow amount is reduced by melting. The snow melting is parameterized by

$$\mathrm{M} = \mathrm{K} \times (T - Tcrit). \qquad (20)$$

for T>Tcrit. The melting coefficient (K) is chosen to be *K*=90 *mm/month/°C* (Braithwaite, 1995). Thus a month with an air temperature of 10°C will remove 900 *mm* snow and put the equivalent amount of water into the soil. Note that snowmelt is not characterized as a time scale, as the rate is not proportional to the amount of snow. Melting snow and rain are added to soil moisture.

The assumption that snow melt is proportional to air temperature has been defended by Ohmura (2000). He argues that the two largest heat sources for snowmelt, sensible heat and atmospheric longwave radiation, are both characterized by air temperature.

The above equations are used to compute the soil moisture and snow remaining at each grid point at the end of each month. For a climatological calculation, the calculation is done sequentially, month by month, over a period of two years. In the first year, (i.e. the "spinup" year) we start with dry conditions at the end of September. This is a valid assumption in most dry areas. In the second year, with a September starting value computed from the spinup year, the results are independent of the starting values everywhere, even in the moist areas.

Eq (9) is simple enough to allow the derivation of analytical solutions describing how soil moisture responds to simple inputs. For example, if $T(t)$ and $PET(t)$ are constant throughout the year and $P(t)$ is sinusoidal, given by

$$P(t) = \hat{P} \times \left(1 + e^{i\omega t}\right) \qquad (21)$$

The soil moisture is

$$SM(t) = \hat{P}\left(\frac{F - i\omega}{F^2 + \omega^2}\, e^{i\omega t} + \frac{1}{F}\right) \qquad (22)$$

where $i = (-1)^{1/2}$, $\omega = 2\times\pi/12$ and F is the inverse of the soil moisture storage time.

For example, if P_{tot}= 240 *mm* (so $\hat{P}$ =20 *mm/month*), *MSM* = 100 *mm, PET* = 50 *mm/month and Sz=0* then *F*=0.5 and *SM*(*t*) varies between 68 and 12 *mm* with the time of maximum soil moisture delayed 1.5 months from the time of maximum precipitation. As *SM* stays below *MSM* in this example, no runoff would be generated. The actual *SM* profiles computed from the model differ from this ideal case as the seasonal temperature cycle is included and the snow and runoff thresholds are hit in the winter in the colder wetter areas, allowing snow accumulation, soil saturation and excess runoff.

The hydrology model can be summarized by listing its coefficients, as in the Table below

symbol	*name*	*value*	*units*
$\sum W$	Sum of weights	1.1	none
γ	Lapse rate	-5	C/km
C	PET/Temperature slope	5.2	Mm/month*C
α	Runoff/roughness slope	0.02 or 0.05	$(\text{month*meter})^{-1}$
K	Snowmelt rate	20 (or 90)	Mm/month*C
MSM	Max Soil Moisture	100	mm
Tcrit	Intercept temperature for evap. and melting	-2.5	C

While the input climate data provides the seasonal cycle and smooth spatial gradients in temperature and precipitation, the input topography is responsible for most of the small scale structure seen in the computed soil moisture and runoff patterns. It is important to be aware of which physical processes the model is invoking to produce these patterns. First, the model lapse-rate correction decreases the temperature with altitude, decreasing the potential evaporation and, in the highest terrain, allowing precipitation to be stored for a few months as snow. Second, the roughness decreases the soil moisture storage time by increasing the runoff rate. These two factors tend to compensate each other to keep the soil moisture rather low. In high rough terrain, water is removed quickly by runoff while in low warm climates, water is removed quickly by evaporation. An exception would be in a high flat cool basin, where significant amounts of soil water could be stored for several months.

It is equally important to know what physical processes are not included in the model. We have not included enhanced precipitation with height, soil variations (except as correlated with roughness), collection of runoff in basins or modified evaporation with altitude , slope or aspect. The influence of vegetation has been ignored.

When this hydrology model is applied to southwest asia, several patterns are seen which illustrate both the hydrology of the region and the characteristics of the model. Generally, high mountain regions have little soil moisture, even in a wet season like March. The highest mountains have snow, but the soil moisture is low due to rapid runoff and the lack of melting. Examples include Suphan Dag (4434m) and Mt Ararat (5165m) in Turkey and Kuh-e Sahand (3722m) in Iran. Lower mountains in the desert and steppe areas are dry due to small precipitation and moderately rapid runoff. Examples are Jebel al Aziz in northern Syria and the antilebanon just west of Damascus. The wettest regions are high mountain basins such as Lake Van in eastern Turkey and Lake Urmia in northwestern Iran. There, precipitation is significant, while temperature and roughness are low. Large areas of moderate soil moisture are seen in the arc of the fertile crescent including Al Jazirah and Mesopotamia where moderate precipitation is matched by low roughness. In the deserts of Syria, Jordan, Iraq and Saudia Arabia, several flat basins have low but significant soil moisture, in spite low precipitation and high temperature, due to their extremely low roughness(e.g. Sz= 2m). These basins include Palmyra and Damascus. Our model does not include groundwater or river flow into these basins, but it does reduce runoff to a negligible level, thus allowing them to hold more moisture than the nearby hills.

Figure Captions

1. Map of the Middle East study area with terrain, national boundaries, watershed boundaries and WMO climate stations.
2. Temperature classes for the Middle East. Each 5km pixel is assigned to one of nine classes based on its seasonal cycle of temperature. Class signatures are shown in Fig. 3
3. Seasonal cycles for each climate class. Each point represents an average value for the class: (a) temperature classes from Fig 2; (b) precipitation classes from Fig.3; (c) soil moisture classes from Fig 4.
4. Precipitation classes for the Middle East. Each 5km pixel is assigned to one of ten classes based on its seasonal cycle of precipitation. Class signatures are shown in Fig 3.
5. Soil moisture classes for the Middle East. Each 5km pixel is assigned to one of ten classes based on its seasonal cycle of soil moisture, computed from the hydrology model. Class signatures are shown in Fig 3.
6. Scatterplots of relationships between inputs and outputs of the hydrologic model applied to the Middle East. Each point represents a gridpoint in the model domain (subsampled) (a) Length of the soil moisture season (days) versus the annual precipitation (cm); (b) Number of days with sufficient soil moisture versus the KGMI computed from (1); (c) Annual runoff versus annual precipitation.
7. Vegetation classes for the Middle East (SWAP13/93). Each 1km pixel is assigned to one of thirteen classes based on its seasonal cycle of NDVI. Satellite data is 10-day Composite AVHRR for February to October 1993. Class signatures are shown in Figure 8. Class legends are in Table 2. Contours of Wetdays from the hydrology model are overlain in solid black lines.
8. Class signatures for SWAP13/93.Each point represents an average value for the class for the 10-day compositing interval. (a) classes 1 though 7; (b) classes 8 through 13.
9. January snow frequency in Anatolia for 1983 to 1994; derived from 8km composite AVHRR data. Snow is detected using channels 2 and 3?. Color indicates the number of years in the interval snow was present.
10. Snow cover time series. The winter cycle of snow cover for Anatolia is plotted for each year. Note the magnitude of the difference between 1993 and 1985. The 12-year average cycle is also shown, along with three hydrology model predictions: control, 5°C warmer and 5°C colder.
11. Comparison of actual and predicted snow cover in Anatolia for January. Green indicates that snow was observed but not predicted. Yellow indicates that snow was predicted and observed. Red indicates that snow was predicted but not observed. Prediction is based on threshold monthly temperatures (-2.5°C) with a lapse rate correction. Observed snow implies that snow was seen in at least half of the years between 1983 and 1994.
12. Regression lines for Anatolian snow cover versus mean DJF temperature anomoly. Three sets of climate stations are used: T1,T2 and T3 corresponding to western, central and eastern Anatolia.
13. The non-cumulative histogram in the Anatolian test region. Abscissa is the altitude above sea level. Ordinate is the surface area (km^2) above the given altitude. A linear fit is shown (5).
14. River discharge profiles for the Euphrates, Upper Tigris and Greater Zab; computed from the 5-km gridded hydrology model using climatological monthly temperature and precipitation. Observations from pre-dam periods are included from gauging stations at Hit, Mosul and Girdmamukk. In addition to the reference run, curves are shown for runs with perturbed climates and maximum soil moisture.

15. Irrigated areas, as they appear in a large scale 1km AVHRR NDVI image from August. Selected small regions are outlined. (a) Syria and southeastern Turkey; (b) Mesopotamia
16. Change of irrigation patterns in Mesopotamia. The difference between irrigated area per pixel (from (16)) in August 1983 and 1993 is shown in false color. Data is the 8-km composite AVHRR.
17. Change of irrigation patterns in three small regions: (a) east of Aleppo, Syria, (b) Harran Plain, (c) near Kirkuk, Iraq. Irrigated area per pixel from (16) is shown. Data is the 1-km composite AVHRR.
18. Change of irrigation patterns near the junction of the Khabur and Euphrates Rivers. The difference between NDVI in August 1984 and 1990 is shown in false color. (a) base map, (b) subregion 1, (c) subregion 2, (d) subregion 3. Data is from Landsat TM.
19. Landsat TM zoom of irrigated plantation east of Aleppo. (a) Three channel image (432-RGB), (b) Identified fields
20. Field size spectrum. The base 10 log of the number of fields per size range is plotted against the base 10 log of the field size (hectares). Linear regression is shown.
21. Irrigated area versus time from 1982 to 1998; derived using the proportional method (16). The period 1982 to 1994 is based on 8-km composite AVHRR data. The period 1995 to 1998 is based on individual 1-km AVHRR images. (a) Turkey, (b) Turkey/Syria border and east of Aleppo, (c) along the Euphrates in Syria, (d) along the Khabur in Syria, (e) near Kirkuk and Baghdad, Syria, (f) south of Baghdad and near Basra
22. Climate indices from 1982 to 1994 for a group of WMO stations in northwestern Syria: (a) Mean Dec-Jan-Feb precipitation, (b) mean Dec-Jan-Feb mimimum temperature, (c) mean Feb-Mar-Apr precipitation. These indices are the basis for the correlation maps in Fig 23-25.
23. Sensitivity of spring vegetation to interannual climate variation. Maps of correlation coefficient between NDVI and three climate indices shown in Figure 22. Satellite data is 8-km composite AVHRR from 1982 to 1994. (a) April, (b) May, (c) June.