Evidence for decadal variation in global terrestrial evapotranspiration between 1982 and 2002: 2. Results

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[1] Terrestrial evapotranspiration (ET) cools the surface and moistens the atmosphere near the Earth's surface. Variations in this important climate factor have major environmental and socioeconomic impacts. How terrestrial ET has varied in the past and what caused the variations, however, have remained quite uncertain. These issues are addressed by calculating monthly global ET from 1982 to 2002 at 1120 globally distributed stations, using a modified Penman-Monteith method that was developed in the first part of the two-part paper. Our analyses show that ET has a significant decadal variation (~10%) regionally and globally. Over the period analyzed ET for global land increased by 0.6 W m^{-2} per decade equal to 1.2 W m^{-2} (about 2.2% in relative value) or 15 mm yr^{-1} in water flux during the study period. We show that long-term variations of ET in humid areas such as the tropics, Europe, and humid areas of Asia are primarily controlled by variations in incident solar radiation R_s connected to changes in cloudiness and aerosols. However, soil water supply, estimated here by RH, and connected to precipitation, is the dominant factor in controlling long-term variations of ET in arid areas. A correlation analysis demonstrates that the dependence of ET on R_s switches to negative in dry regions. Furthermore, its dependence on relative humidity switches from negative in moist regions to positive in dry regions. Its dependence on normalized difference vegetation index is uniformly positive.

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1. Introduction

[2] Evapotranspiration (ET) is the sum of soil evaporation, vegetation transpiration, and evaporation from the canopy of intercepted rain, fog and dew. These water transfer processes depend on surface water and energy availabilities as well as atmospheric evaporative demand. ET is influenced by numerous factors, including root depth moisture, surface moisture, stomatal conductance (influenced by leaf area and vegetation coverage), aerodynamic conductance (influenced by wind speed and near-surface air stability), incident solar and longwave radiation flux, air temperature (T_a) and water vapor pressure deficit (VPD). A further complication is that the relationships between ET and the above parameters depend on climate conditions. For example, in arid areas, moisture supply is the dominant parameter but it is less important in humid regions [*Nemani et al.*, 2003; *Teuling et al.* 2009]. In cold regions, T_a is one of the dominant parameters [*Nemani et al.*, 2003].

[3] ET is a major component of the terrestrial water, energy, and biogeochemical cycles. Its regional and global variability and long-term trends are not yet well characterized [*Trenberth et al.*, 2007]. It has been obtained from various satellite remote sensing approaches, but so far only for short durations and with questionable accuracy [*National Research Council*, 2007]. Local site measurements are more accurate but also of short duration.

[4] Various other hydrological measurements are made that may be correlated to long-term ET, such as the evaporative loss from a pan of water. *Peterson et al.* [1995] and numerous later studies [*Lawrimore and Peterson*, 2000; *Golubev et al.*, 2001; *Hobbins et al.*, 2004; *Roderick and Farquhar*, 2004] have reported decreasing pan evaporation over large areas and in different regions of the world before the 1990s. However, pan evaporation does not provide a direct estimate of ET [*Lawrimore and Peterson*, 2000; *Hobbins et al.*, 2004; *Brutsaert and Parlange*, 1998; *Brutsaert*, 2006; *Kahler and Brutsaert*, 2006]. Changes in ET can substantially differ from those of pan evaporation [*Roderick and Farquhar*, 2002; *Wild et al.*, 2004; *Roderick et al.*, 2009]. They have been used with the "Bouchet complimentary relationship" where pan data are available to estimate trends of ET [e.g., *Szilagyi*, 2001; *Hobbins et al.*, 2004], that approximates ET by twice the "potential evaporation" minus the pan evaporation.

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[5] Long-term data records for precipitation and runoff have also been used to constrain ET as a residual. ET derived from this method may have a large uncertainty because of accumulated errors [*Sheffield et al.*, 2009]. Continental precipitation records are available from several data sets. IPCC AR4 indicates that six such data sets provide a variety of trend estimates [*Trenberth et al.*, 2007] for annual precipitation, ranging from −4 to 16 mm per decade over the period 1980–2004 (essentially the period of this study) with no trend estimate being significant. However, intensities of precipitation may have increased and dry spells have become longer. Such changes in intensity and frequency could increase runoff and decrease ET even if the total amount of precipitation were not to change [*Trenberth et al.*, 2007], as supported by *Dai et al.* [2009] showing that global runoff had a significant increase in the period of this study. Another recent but less thorough estimate [*Piao et al.*, 2007] suggested a decrease in runoff. Useful estimates for individual river basins appear possible, but confidence in global trends is precluded by the large spatial variability of precipitation and runoff that makes even the sign of their change questionable.

[6] A more direct approach is developed in the first part [*Wang et al.*, 2010] of the present two paper series; that is, it employs the currently available ET measurements to calibrate a semiempirical formula. That formula is used here with long-term data sets provided by conventional meteorological and weather satellite observations to estimate ET globally for several decades. Our method is simple, has a low sensitivity to errors in input data, and uses data that are easy to obtain; that is, it uses surface incident solar radiation (R_s), T_a, relative humidity (RH), wind speed (WS), satellite-derived normalized difference vegetation index (NDVI), and VPD, a function of T_a and RH.

2. Method

[7] Detailed information about the method has been given in the first part [*Wang et al.*, 2010] of this two-paper series and is summarized here.

[8] The traditional Penman and Penman-Monteith equations [*Penman*, 1948; *Monteith*, 1965] divide ET into an energy controlled (ET_E) and atmospheric controlled (ET_A) part. The ET_E term is approximated as

$$ET_E = \frac{\Delta}{\Delta + \gamma} \cdot R_s \cdot [a_1 + a_2 \cdot NDVI + RHD \cdot (a_3 + a_4 \cdot NDVI)], \quad (1)$$

where Δ is the derivative of the saturated vapor pressure with respect to T_a, γ is the psychrometric constant, and R_s is incident solar radiation. The atmospheric control part ET_A is

$$ET_A = \frac{\gamma}{\Delta + \gamma} \cdot WS \cdot VPD \cdot [a_5 + RHD \cdot (a_6 + a_7 NDVI)]. \quad (2)$$

The total *ET* has the form

$$ET = a_8 \cdot (ET_E + ET_A) + a_9 \cdot (ET_E + ET_A)^2. \quad (3)$$

[9] The nonlinear form of equation (3) is selected primarily because of the issue of saturation of NDVI [*Huete et al.*, 2002]; that is, it loses sensitivity to vegetation coverage (or leaf area) when vegetation is dense. The Enhanced Vegetation Index (EVI) which is available from new generation satellite data, such as the Moderate Resolution Imaging Spectroradiometer (MODIS), is designed to effectively correct the saturation effect of NDVI [*Huete et al.*, 2002] and does not appear to need the nonlinear term; that is, the first part [*Wang et al.*, 2010] of the two-part paper shows that the nonlinear item (a_9) is negligible for EVI and so equation (3) is more like the traditional Penman-Monteith Equation ($ET = ET_A + ET_E$).

[10] The term $RHD = 1 - RH/100$ is used to estimate the control of water supply on ET because global reliable soil moisture data are not yet available. The $\Delta/(\Delta + \gamma)$ and $\gamma/(\gamma + \Delta)$ factors in equation (1) and (2) depend solely on T_a [*Wang et al.*, 2006], and provide, along with R_s, the well known strong correlation of ET with temperature [*Priestley and Taylor*, 1972]. Variations in stomatal conductance are assumed to be included in the VIs and RHD, and aerodynamic conductance is the correlation of ET with the assumed linear function of WS.

[11] The net radiation (R_n), used in the traditional Penman-Monteith method to parameterize ET, is not widely available. However, its variability is adequately captured by R_s and other measured factors in equations (1)–(3). In particular, variations in T_a and atmospheric water vapor content, especially in the near surface layer, control the downward longwave and the cooling effect of ET affects the surface longwave emission. In other words, R_n can be estimated from R_s with the addition of T_a, RH, and NDVI data for various land cover types and different surface elevations ranging from 98 m to 4700 m [*Wang and Liang*, 2009]. All these factors are included in our ET parameterization.

[12] The coefficients were derived by regression using 16 day periods over 5 years at 15 stations selected for the quality of their data and parameter coverage. They were validated with ground-based measurements at 49 additional stations, including the long-term measurements collected by Ameriflux, Asiaflux, the Atmospheric Radiation Measurement (ARM) sites, and additional sites operated by individual principal investigators released by the FLUXNET website. The sites are mainly located in North America and Asia, with the exception of three sites in Australia, two in Europe, and one in Africa. The climates of the sites vary from tropical to subarctic, and arid to humid. Their elevation ranges from near sea surface level to more than three thousand meters above sea level. The land cover types of the sites include desert grasslands, rainfed and irrigated croplands, grazed and ungrazed grasslands, savanna, shrub land, deciduous forest, evergreen forest and mixed forests.

[13] The same model parameters a_1 through a_9 in equation (1)–(3) are used for all the 64 sites including both for calibration and validation. The 16 day daily ET was shown to be estimated with an error (standard deviation) of 17 W m^{-2} (25% in relative value), and with an averaged correlation coefficient of 0.94. The method improves estimates of the spatial variation in ET over previous methods. It is also satisfactory in reproducing the interannual variability at sites with 5 years of data, in both humid and arid regions. The correlation coefficient between measured and predicted annual ET anomalies is 0.85.

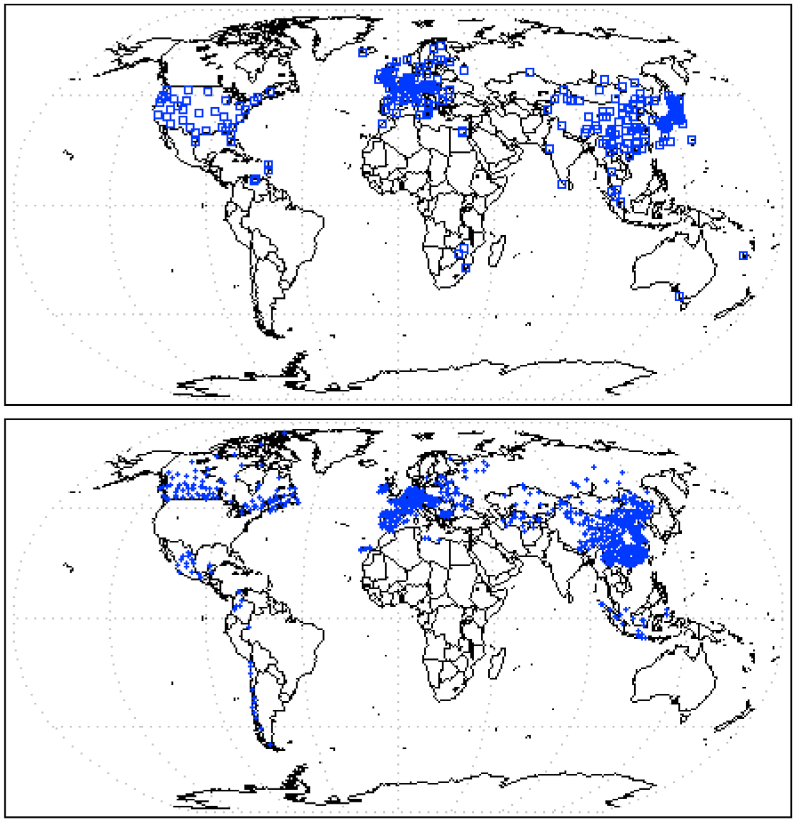

Figure 1. A map of the stations used in this study: (top) stations of Global Energy Balance Archive (GEBA) and (bottom) the stations where sunshine duration (SunDu) data are available. SunDu is used to estimate R_s.

[14] The use of the method on hourly time scales is not advised because in this study aerodynamics conductance is parameterized as a function of WS [*Shuttleworth*, 2007] and ignores the impact of near surface air stability on the aerodynamics conductance.

3. Data

[15] R_s is obtained from direct measurements collected at 270 stations of the Global Energy Balance Archive (GEBA) [*Wild et al.*, 2005] and an empirical formula that uses Sunshine Duration (SunDu) at 850 stations, to accurately estimate the long-term variation of R_s (K. Wang et al., Atmospheric controls on climate variability of surface incident solar radiation, submitted to *Proceedings of the National Academy of Science*, *U.S.A.*, 2010). Additionally, NDVI data sets derived from Advanced Very High Resolution Radiometer (AVHRR) acquisitions at a spatial resolution of 8 km [*Tucker et al.*, 2005], and meteorological observations from the Integrated Surface Hourly Database (ISH) released by National Climate Data Center (NCDC) are used. Because the R_s measurements from GEBA are only available monthly, we first averaged all the other data sets to monthly values, including R_s derived from SunDu, RH, WS, NDVI, VPD and T_a. Data were selected that cover the time period from 1982 to 2002 and to be concurrent for at least 168 months during the study period. These 1120 stations, as shown in Figure 1, are globally distributed, but more densely in Europe, Asia and North America.

[16] Estimates are sparse for Africa and Australia where there are few or no ground-based measurements of R_s (GEBA) or reported values of SunDu for the period studied. This study only uses the stations that reported SunDu to the WMO (World Meteorological Organization). Long-term SunDu measurements in Australia [*Lamb et al.*, 2009; *Suehrcke*, 2000] and Africa [*El-Metwally*, 2005; *Chineke*, 2008] are not readily available.

[17] *Pinker et al.* [2005] analyzed satellite data from 1983 to 2001 and found on average a slightly negative trend in R_s over land. Their data set provides better global coverage but does not consider the long-term variations of tropospheric aerosols over the past 3 decades [*Wang et al.*, 2009]. Substantial discrepancies exist between satellite-derived data and pyranometer measurements because of the strong effects of aerosols on surface shortwave irradiance [*Hayasaka et al.*, 2006; *Xia et al.*, 2006; *Wild*, 2009]. Furthermore, the ISCCP cloudiness used to estimate R_s [*Pinker et al.*, 2005] may be affected by satellite viewing geometry artifacts that are not related to physical changes in the atmosphere. Consequently, it may be inappropriate to use this data for trend estimation of long-term variability [*Dai et al.*, 2006; *Evan et al.*, 2007].

[18] NDVI is another important factor determining the long-term variation of ET at regional scale. The influences of volcanic stratospheric aerosols, solar zenith angle and sensor degradation on NDVI have been corrected, and AVHRR NDVI is consistent with NDVIs from other satel-

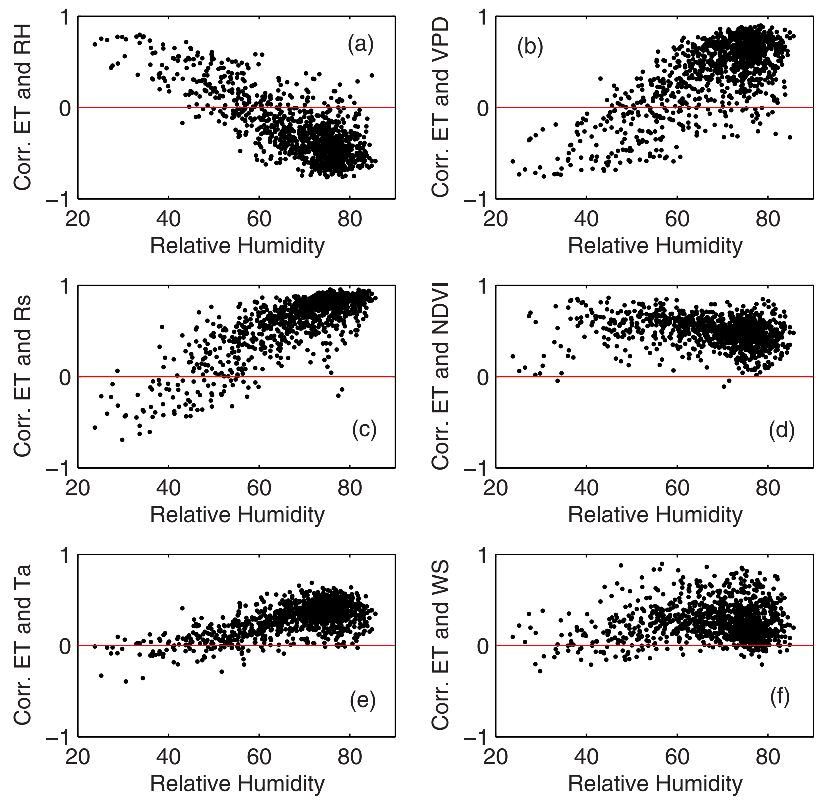

Figure 2. Scatterplots of the correlation coefficients between *ET* and (a) RH, (b) VPD, (c) R_s, (d) NDVI, (e) T_a and (f) WS as a function of multiyear averaged RH. The correlation coefficients are calculated using the monthly anomalies from 1982 to 2002 at each station. Each point in the plots represents one station.

lite sensors [*Tucker et al.*, 2005]. The data set used in this study is an improved version [*Tucker et al.*, 2005] of the NDVI data set previously used to study the long-term variation of terrestrial vegetation [*Myneni et al.*, 1997; *Lucht et al.*, 2002; *Nemani et al.*, 2003].

4. Results

[19] Monthly anomalies of ETs are derived by removing the seasonal cycle averaged from 1982 to 2002 at each station. Figure 2 shows as scatterplots the correlation coefficients between ET and R_s, NDVI, RH, VPD, T_a and WS. These were calculated using the monthly anomalies from 1982 to 2002 at each station and are shown as a function of multiyear averaged RH; high in humid regions and low in arid regions. Figure 2a shows that for low values of averaged RH, its fluctuations are positively correlated with ET, but at high RH, they are negatively correlated. For dry conditions, ET contributes to RH, but under moist conditions VPD drives ET. The correlations between ET and VPD are similar in magnitude to those with RH but opposite in sign as expected from their definitions: $VPD = e_s^* (1 - RH/100)$, where the atmospheric saturation water vapor pressure is a function of T_a.

[20] Figures 2c and 3e show that R_s has the strongest correlation with ET in humid areas and that the coefficients are less where it is drier, becoming negative in arid areas. NDVI is positively correlated with ET everywhere and overall it has the second largest such correlation (Figures 2d and 3d). The association of NDVI with ET generally increases for drier conditions; that is, vegetation's variability with ET is strongest in semiarid areas, likely because both covary with precipitation (Figures 3f and 4f in 1990s). For desert conditions, vegetation is spare and contributes little to ET.

[21] Figure 2e shows that T_a and ET are strongly correlated in humid areas but less so for drier conditions and their correlation becomes zero or negative in semiarid or arid areas. Similarly, Figure 2f shows that ET is most strongly affected by WS in areas of moderate to high humidity. This variable determines the aerodynamic conductance between the surface and the lowest layer of the atmosphere, a term comparable in its effect to stomatal conductance except for tall canopies [*Sellers et al.*, 1997].

[22] Figure 3 shows the average of the station data over various regions, Europe, North America, Asia, the Southern Hemisphere (where there are relatively few stations), the tropics (20°N–20°S), and in arid regions defined by multiyear averaged RH is less than 50%. Table 1 summarizes the correlation over these regions between ET anomalies and the anomalies of other parameters. Figure 3 compares the variation of ET with that of R_s. The decadal variations of ET are typically ~5 W m^{-2} and largely follow those of R_s, tracking it best in tropical regions (Figure 3e). ET shows large departures from R_s in more arid regions, where its variability is less pronounced than that of R_s. In particular Figure 3f shows a big dip around 1990 in arid areas associated with a period of relatively high precipitation in northern Asia [*Wang et al.*, 2003] that also shows up in the

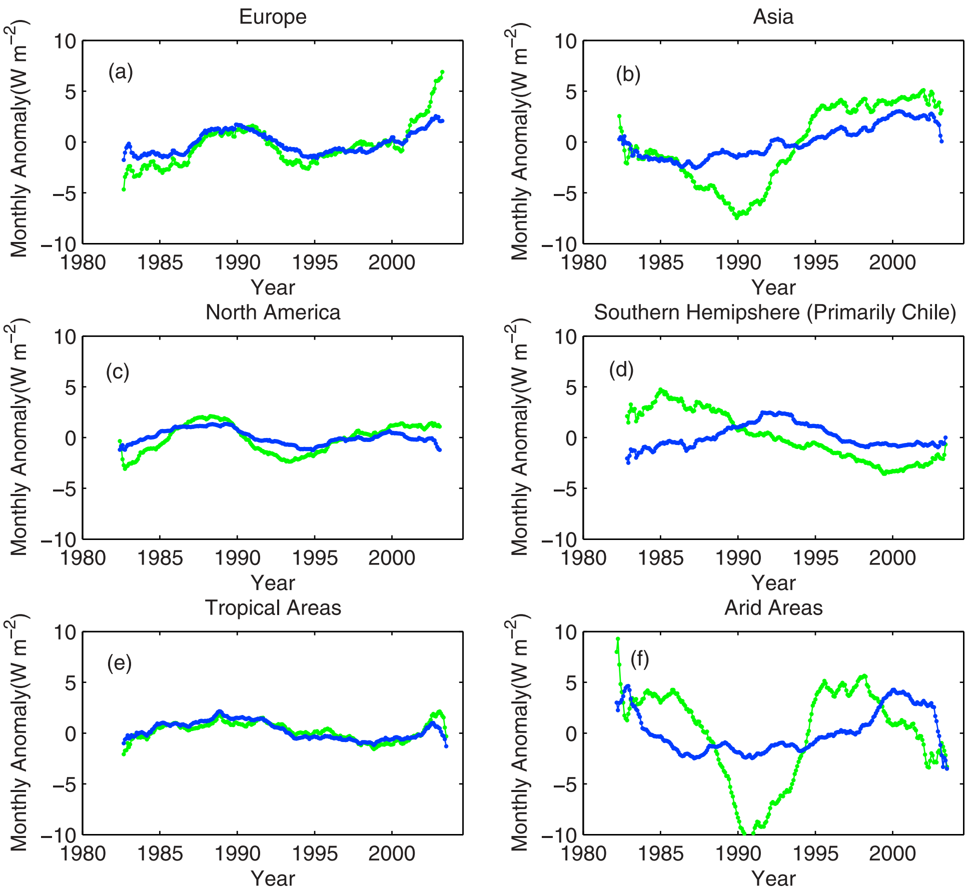

Figure 3. The times series of 5 year smoothed ET (blue line) and R_s (green line) anomalies for (a) Europe, (b) Asia, (c) North America, (d) the Southern Hemisphere, (e) tropical areas, and (f) arid areas. Tropical areas refer to regions where latitude is between 20°S–20°N, and arid areas refer to regions where multiyear average relative humidity is less than 50%. ET and R_s demonstrate significant decadal variation in all the regions.

Asia average. After the dip, starting in 1995, the ET in arid areas shows a rapid increase up to 2000 although R_s is flat or declining. Figure 4 shows RH and NDVI as two other factors in addition to R_s important for determining ET. The variability of RH is seen to account for the larger departure of ET from R_s seen in Figure 3. In particular, it shows that the arid area has near constancy of ET from 1985 to 1995, but a ramp up after 1995 in spite of the rapid increase of R_s from 1990 to 1995 and its constant or declining values from 1995. Likewise, the rapid "brightening" that occurred in North America and Europe since 1995 has a weaker effect on ET because of the drier conditions (i.e., a decreasing RH). Figure 5 shows the wind anomalies by region. Evidently, the contribution of this term to the variability of ET is confined to the Southern Hemisphere data set (primarily Chile).

[23] Table 1 summarizes the correlations by regions between ET anomalies and the anomalies of other parameters by region. For the most part it shows that the correlations with ET seen in Figure 2 are maintained, region by region and that the correlation from R_s is strongest outside of arid regions, where it flip signs to a small negative. The correlations with RH and VPD are of opposite sign, as expected, except for arid regions. Thus, the increase of RH with ET as expected from wet weather also leads to small increase of VPD. Such a result could be expected in areas with strong temperature variability causing changes of VPD to be of the same sign as vapor pressure and provided warmer weather were also wetter, leading to more RH.

[24] Estimates of the linear station trends from the anomalies are obtained using the Mann-Kendall trend test at each station. About 44% of the stations pass the 95% confidence test. The trends may be seriously impacted by the beginning and ending date of the period, which can vary for different stations. To reduce the influence of heterogeneity of the station distribution, global averages of long-term trends of ET and other parameters are obtained by aggregating the trends at the stations into 5° × 5° grid boxes, yielding 266 such boxes total. The global trend of daily R_s is 1.0 W m^{-2} per decade averaged from over the 266 grid boxes. The trend in global WS is −0.2 m s^{-1} per decade; that is, WS had a ~5% reduction during the study period. The long-term trends of RH and NDVI are very small. VPD increased at a rate of 0.04 hPa per decade (global average of VPD has a magnitude of 10 hPa) mainly from global warming. As a result of the increase in R_s and VPD, and decrease of WS, daily global ET increased by 0.6 W m^{-2} per

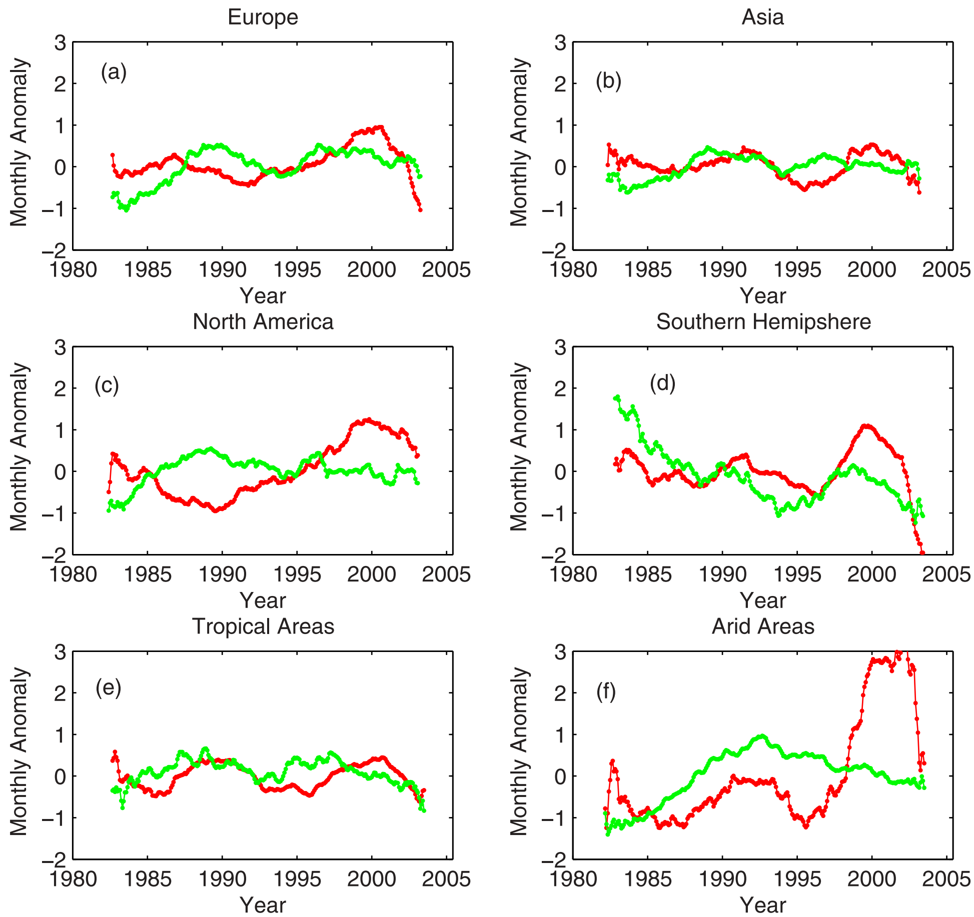

Figure 4. The times series of 5 year smoothed RH (red) and NDVI × 100 (green) anomalies for (a) Europe, (b) Asia, (c) North America, (d) the Southern Hemisphere, (e) tropical areas and (f) arid areas. Tropical areas and arid areas are defined in Figure 3.

decade, equal to 1.2 W m^{-2} (about 2.2% in relative value for a global averaged ET of about 55 W m^{-2}) or 15 mm yr^{-1} in water flux over the entire period.

5. Discussion

[25] This analysis shows how ET has changed at regional or continental scale for the period 1982 to 2002. These estimates are indirect, depending on a semiempirical expression, tailored in structure to the formula of Penman Monteith, but formulated to make use of standard meteorological data.

[26] The first part of the two-series paper shows that the variability in time over the sites with 5 years of data is adequately captured [*Wang et al.*, 2010], indicating that this variability largely corresponds to that of cloudiness and aerosols. However, because the variability of the data used to derive our semiempirical expression for ET was mostly spatial and seasonal, its estimation of multidecadal trends may be biased, especially if affected by other long-term trends that are not represented in the data, e.g., the stomatal closure from increasing CO_2 [*Gedney et al.*, 2006; *Betts et al.*, 2007], a factor that reduces ET by about 0.05 mm yr^{-1} but may be largely or entirely compensated by increases in leaf area [*Piao et al.*, 2007].

[27] What parameters control ET is implicit in the expression, equation (2), used to determine ET. However, their relative contributions are not obvious, because of the nonlinear dependence of ET on them. In particular, it is negatively dependent on RHD especially for sparse vegetation; that is, a_3 and a_6 are negative but the atmosphere control term is less affected, i.e., $|a_6|/a_5$ smaller than $|a_2|a_1$, and is further enhanced by its positive dependence on VPD. Figure 2 and Table 1 quantify the relative dependence of ET on the other climatic factors determining it. They show that the long-term variation of R_s is the dominant factor controlling the long-term variation of ET at regional scales in humid regions, including the tropics, Europe and humid regions in Asia.

[28] The increase of solar radiation in Europe is consistent with long-term decreases in atmospheric aerosols there [*Norris and Wild*, 2007]. Although aerosols substantially increased in China and India [*Wang et al.*, 2009], R_s increased [*Shi et al.*, 2008] in middle and north China, apparently because of decreasing cloud cover [*Qian et al.*, 2006] (Figure 6), while R_s decreased by −0.17 W m^{-2} to −1.44 W m^{-2} from 1981 to 2004 at twelve stations in India [*Padma Kumari et al.*, 2007]. Aerosols have increased over South America, Australia and Africa [*Wang et al.*, 2009] but clouds have decreased over these regions (see *Warren et al.* [2007] and Figure 6). Therefore, it is problematic to conclude

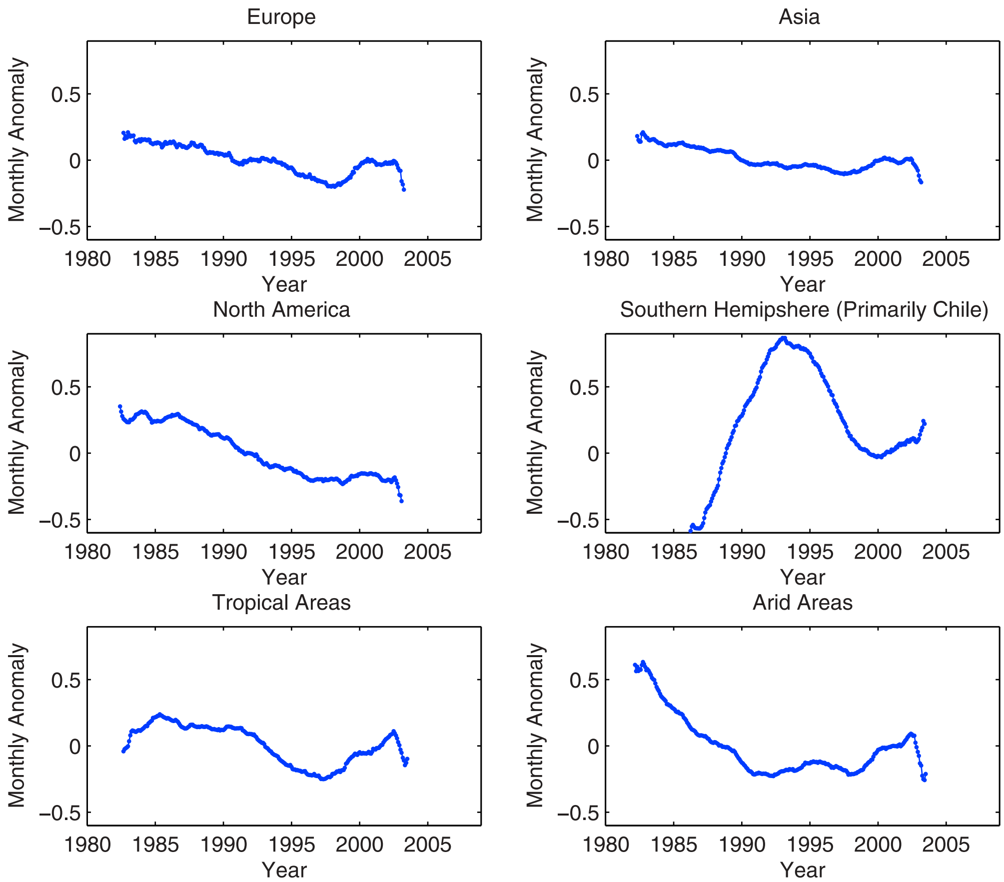

Figure 5. The times series of 5 year smoothed WS anomalies for (a) Europe, (b) Asia, (c) North America, (d) the Southern Hemisphere, (e) tropical areas, and (f) arid areas. Tropical areas and arid areas are defined in Figure 3. Wind speed substantially decreased in Europe, Asia, North America, and arid area.

whether R_s has increased or decreased in the Southern Hemisphere during the study period.

[29] The dominant control of R_s on ET in energy-limited regions has been implicit in many previous expressions developed to estimate ET. In particular, *Priestley and Taylor* [1972] assumed ET to be linearly proportional to R_n with air temperature-dependent coefficients. However, other authors have included a VPD-dependent atmospheric control, in particular Penman and Monteith. Figures 2a and 2b show that this is also a significant control for highly humid regions. Because seasonal variability has been removed, the variability of R_s will be connected to that of RH and VPD largely through their mutual connections to cloudiness. Thus, at high humidity, some of inferred dependence of ET on VPD could be a dependence on R_s or vice versa, for example, as might be seen by some other form of fitting such as by linear multiple regression or some different nonlinear formula. The presence of these correlations suggests why the simple Priestley Taylor expression has been reasonably successful when net radiation is available.

[30] The different structure of the dependences at low RH is not surprising but, to our knowledge, has not been previously captured by such a semiempirical approach as this. It is well established that soil moisture, hence ET, is limited by the supply of water under dry conditions. Budyko related ET to atmospheric forcing as given by R_s and precipitation P [e.g., *Koster and Suarez*, 1999]. However, the water control on ET is more directly through soil moisture than P, excluding interception losses [e.g., *Emanuel et al.*, 2007]. RH increases in dry climates with occurrence of P, and through its

Table 1. Correlation Coefficients Between Anomalies of ET, R_s, NDVI, RH, WS, and T_a at Regional Scales in Six Subregions[a]

	Europe	Asia	North America	South Hemisphere	Tropical Areas	Arid Areas
ET, R_s	0.86	0.59	0.58	0.59	0.80	−0.12
ET, VPD	0.74	0.43	0.52	0.06	0.26	0.17
ET, RH	−0.54	−0.07	−0.23	−0.06	−0.20	0.62
ET, NDVI	0.48	0.69	0.58	0.44	0.64	0.40
ET, WS	0.01	0.07	0.16	0.45	0.44	−0.34
ET, T_a	0.57	0.50	0.43	0.11	0.21	0.02

[a]Here tropical areas refers to the Areas where latitude is between 20°S and 20°N. Arid areas refers to the region where multiyear average relative humidity is less than 50%.

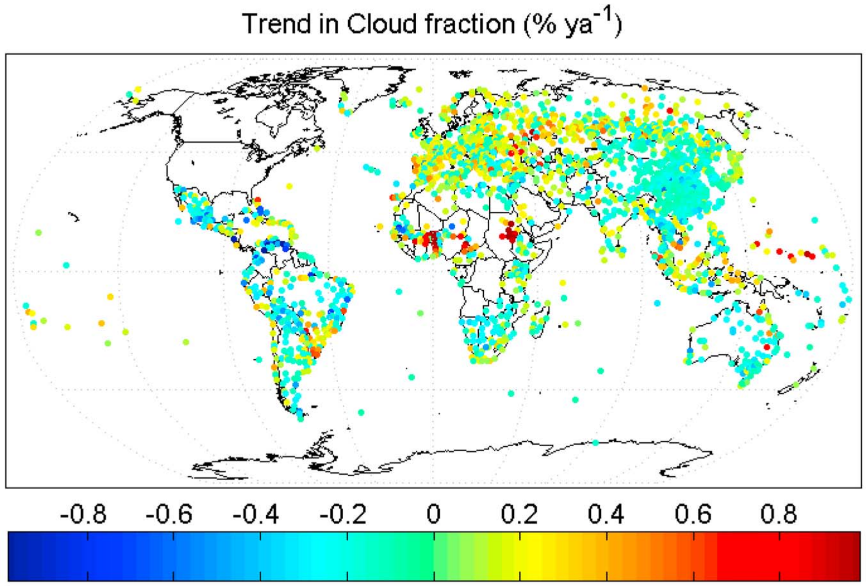

Figure 6. Linear trend of daily total cloud coverage from 1973 to 2008. The monthly total cloud cover fraction anomaly is derived and used to calculate the linear trend using the Mann-Kendall trend test method, and only stations that pass the 95% significance level in the Mann-Kendall trend test are shown. Some sites over North America and some European countries changed the observational method from human visual observations to instrument observations during the 1990s and are excluded because they show obvious discontinuities in total cloud coverage.

soil moisture and ET. Thus, under dry conditions it is as plausible an index for ET as is *P*. However, under moist conditions, an anomaly of VPD; that is, a negative anomaly of RH will drive ET (Figure 2). Figure 2 indicates the crossover point for these two regimes is RH of around 50–55%. It also shows that dependencies on T_a and WS largely disappear in the dry regime. The dependence of ET on NDVI (Figure 2d) appears to be largely independent of RH, but is likely to involve different processes in dry versus wet regimes. Under dry conditions, vegetation will obviously increase in greenness with more precipitation and hence ET. However, under moist conditions, more NDVI indicates more leaves, hence higher canopy conductance, i.e., the stomatal dependence of the Penman-Monteith expression.

[31] During the study period, WS has a large long-term negative trend in many regions [*Xu et al.*, 2006; *Roderick et al.*, 2007] that also contributes to the long-term decrease in water evaporation from a pan [*Roderick et al.*, 2007]. WS appears to have considerably less effect on the variability of ET than it does on pan evaporation [*Roderick et al.*, 2007].

6. Conclusions

[32] This is the second part of the two-part paper. In the first part, we developed a modified Penman-Monteith method that can use simply obtained data to estimate ET [*Wang et al.*, 2010]. This method is simple and has a low sensitivity to errors in input data. This second part is focused on how *ET* varied over the land from 1982 to 2002 and what caused the variations. These issues are addressed by calculating monthly global ET at 1120 globally distributed stations.

[33] The method uses data for R_s, WS, T_a and RH, and some index of vegetation cover, e.g., NDVI. By formulation, the estimates of ET correlate with all these parameters. Relative dependences are not obvious because the method depends on them nonlinearly. A correlation analysis demonstrates that the dependence of ET on R_s is strongest in moist regions, but switches to negative in dry regions. Furthermore, its dependence on relative humidity switches from negative in moist regions to positive in dry regions. The dependence on NDVI is uniformly positive.

[34] Our analyses show that ET has a significant decadal variation (~10%) regionally and globally. This variability is connected to changes in cloudiness and aerosols and thus R_s in moist regions, but more dependent on fluctuations of precipitation in more arid regions. Over the period analyzed ET for global land increased by 0.6 W m^{-2} per decade equal to 1.2 W m^{-2} during the study period (about 2.2% in relative value since global averaged land ET is ~55 W m^{-2}) or 15 mm yr^{-1} in water flux.

[35] We show that long-term variations of ET in humid areas such as the tropics, Europe and humid areas of Asia are primarily controlled by R_s. However, soil water supply, estimated here by RH is the dominant factor in controlling long-term variations of ET in arid areas.

[36] The characteristic distinction between the Penman and Penman-Monteith Equation is that the stomatal conductance i is included in the latter. According to *Jarvis* [1976], stomatal conductance is a function of vegetation characteristics (e.g., leaf area), soil moisture, T_a, and other parameters. In this study, besides T_a, we use NDVI (an index of leaf area) and RHD (1-RH/100) to parameterize stomatal conductance. We use RHD instead of soil moisture in this study because a global reliable soil moisture data set

is not available. Model simulations have shown that RH also works well to quantify the effect of water stress on photosynthesis activity, i.e., the coupling process between ET and carbon uptake [*Mu et al.*, 2007]. The validation with short-term measurements in part 1 of the two-part paper shows that our model works well in predicting seasonal and interannual variations of ET [*Wang et al.*, 2010].

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